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## Acknowledgements

The completion of this MSc thesis demonstrates two years of continuous hard work and dedication, and despite times of stress, lacklustre, and self-doubt the overwhelming enjoyment of studying glaciers spurred me to continue.

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Finally, but most importantly, I admire my parents who supported me both financially and emotionally. This research would have also been impossible without them.

## Declaration

I certify that this thesis consists of my own original work. All quotations from published and unpublished sources are acknowledged as such in the text. Material derived from other sources is also indicated.

Signature Date

# Abstract

High specific discharges from Himalayan headwater basins arises from seasonal winter snow cover, summer monsoon precipitation and melting glacier ice in varying proportions and differing absolute quantities along west –east axes of the Karakoram and western Himalaya. Discharge records for stations in the UIB, Jhelum, Chenab and Sutlej have been examined for periods between 1920 and 2009, together with precipitation and air temperature data for stations with long records (within the period 1876 to 2013) at elevations between 234 and 3015 m a.s.l. Ice-cover age in the basins was between 222.8 and 15,061.7 km$^2$. Climate in the Karakoram had a distinct opposing regime to climate in the western Himalaya. Totals of summer and winter precipitation at climate stations in the UIB reached their maxima in the 1990s. Winter precipitation in the western Himalaya generally declined from the 1980s, while summer precipitation declined since the 1950s but recovered from the 2000s. Mean summer air temperature had an inverse relationship with seasonal totals of precipitation. Air temperatures at UIB climate stations reached their maxima in the 1970s before declining to the 1990s, failing to recover in the 2000s. Air temperatures from mountain stations in the Himalaya displayed maxima in the 2000s, as did climate stations in Tibet.

Riverflows of partially glacierised basins varied regionally, being influenced by regional climate and amount of ice coverage within a basin. Discharge in the glacierised Hunza declined from maxima in the 1970s following a reduction of mean summer air temperature and increasing precipitation since the 1990s. Riverflow in the eastern Karakoram, in the glacial Shyok basin inclined from the 1990s in response to increasing temperatures in Tibet. Annual riverflow along the Sutlej declined from the 1970s-2000s, reduction of flow in the range of 5-31%. Flows at large dams were influenced by riverflows upstream of smaller sub-basins. Annual discharge was least variable from year-to-year at Tarbela (Indus), flow being influenced not only by precipitation fluctuations but also by changes in air temperature affecting glacier melt in headwater basins. Riverflow at Bhakra (Sutlej) also had low variability, flow being influenced by winter precipitation in the upper Sutlej and by summer monsoon in the lower Sutlej. Mangla had high variability of annual runoff, as did other basins with small proportions of ice cover. Basins which had intermediate proportions of ice cover moderated intra-annual variations in runoff as in warm dry years icemelt enhanced riverflows, compensating for reduced precipitation.

# 1. Introduction

Glacierised mountain catchments provide large quantities of runoff from precipitation and seasonal melting of snow and ice to the headwaters of large continental rivers. Usually, total yield of annual discharge in glacierised catchments is in excess of contemporary levels of precipitation as glaciers recede from Little Ice Age maximum extents. Year-to-year riverflow is moderated in glacierised basins by enhanced icemelt in years with low precipitation (e.g. Collins and Taylor 1990). This phenomenon particularly applies for the Upper Indus Basin (UIB) in northern Pakistan where annual totals of precipitation average 223mm $yr^{-1}$. Glaciers and melting of seasonal snow cover in the Karakoram provide critical components of runoff used for irrigation and hydroelectric power, providing a lifeline for populations that depend on these water resources in an otherwise arid area (e.g, Archer et al 2010). In northwest India, however, whilst glaciers do provide an important source of water during the summer melt season, precipitation, particularly monsoon precipitation, provides much additional runoff in lower portions of river basins.

Glaciers in the Himalayas and Karakoram have gained significant attention surrounding their state of health in recent years, but actual mass balances of glaciers in the region remains largely disputed. A report commissioned by the World Bank (Briscoe and Qamar 2007) referring to glaciers in the Indus basin stated; 'It is now clear that climate change is already affecting these western glaciers in dramatic fashion', and, 'there will be fifty years of glacial retreat, during which time riverflows will increase'. Briscoe and Qamar continue to explain; riverflows in the Indus will decrease by a terrifying 30-40% in one hundred year's time as glacial reservoirs dwindle. Such estimates for which the World Bank made this statement were based on future climate warming scenarios by Rees (2005 in Briscoe and Qamar 2007), with earlier evidence of summer cooling from meteorological stations in the UIB being disregarded (e.g, Fowler and Archer 2006). The same controversial statement was repeated in 2009 by the Woodrow Wilson Centre (Kugelman and Hathaway 2009); referring to the Indus basin, they stated: 'Few—if any—areas of the world are suffering from the effects of climate change as much as this legendary mountain region'. On the contrary Hewitt (2005) coined the term 'Karakoram anomaly', presenting largely qualitative evidence of glacier expansion since the 1990s confined to the highest watersheds in the Karakoram. Best estimates from remote sensing also suggests a stable or slight positive mass balance for glaciers in the central Karakoram during the past decade, deviating from glacier decline in the central Himalayas (e.g, Kaab et al 2012). The Karakoram anomaly has been supported indirectly from observations of reduced riverflows in glacierised Karakoram basins (Sharif et al 2013) and from increasing winter and summer precipitation (Archer and Fowler 2004). Growth of glaciers, however, has not been confined to the central Karakoram. Positive mass balance has been inferred in the western Pamir mountain range, giving credence to the 'Pamir-Karakoram anomaly' (Gardelle et al 2013).

In 2007 the IPCC emphasised concerns regarding glaciers in the central Himalayas, suggesting accentuated anthropogenic greenhouse gas emissions had led to accelerated glacier decline, and if the present warming rate continued there would be a high likelihood of them disappearing by the year 2035. The IPCC later retracted their statement in 2010. It is now generally accepted that Himalayan glaciers are receding at rates not dissimilar to glaciers elsewhere in the world (e.g, Bolch 2012), although a recent study (Vincent et al 2013) suggests glaciers of the Lahaul and Spiti region, western Himalaya, were balanced or slightly positive during the 1990s.

Climate controls which affect riverflows and glacier mass balance across the Himalayan arc vary considerably from east to west. From June to September, under the influence of the Indian monsoon, precipitation is drawn from the Bay of Bengal and moves northward producing wetter regions along the indo-Gangetic plain, Bangladesh, and eastern India. Precipitation is enhanced in the western Himalayas in the vicinity of the Sutlej basin (Bookhagen and Burbank 2010) but decreases westward thereafter, rarely penetrating as far north as the Karakoram Mountains. Lower elevations of incised valleys are generally arid in the west but precipitation increases with elevation. In winter and early spring origins of precipitation are of westerly sources and nourish glaciers after summer melting, but decrease rapidly eastward.

Riverflows from Glacierised Himalayan basins generally reflect climatic conditions both annually and seasonally. 80-90% of total annual runoff is derived during the summer season, but the proportion of meltwater runoff varies from east to west, with the Karakoram-Himalaya being largely dominated by snow and icemelt from higher elevations. Riverflows in basins of the humid Greater Himalaya are largely dominated by the monsoon which contributes to runoff at all elevations. Glaciers add a small component to runoff in the east and receive additional mass from snow falling at higher elevations throughout summer. Western Himalayan catchments - Jhelum, Chenab and Sutlej are situated in the transitional zone and receive runoff from melting winter snowpack and direct runoff from the monsoon. Specific runoff tends to increase downstream in Himalayan catchments as precipitation from the monsoon increases along the southern foothills.

## 2. Aims

This study aims to generate a complete survey of long term annual runoff and climate variation for the western-central Himalayan arc, extending from the Indus basin in Pakistan to the Sutlej basin in northwest India. Glaciological relevant variables which affect runoff in glacierised catchments will be identified and their relationship between total annual discharge will be established. Such studies utilising riverflow and meteorological records have been carried out in central Himalayan catchments (Collins et al 2013), and Indus basin (Archer 2003, Khattak et al 2011, Siddique and Hashmi 2012, Sharif et al 2013). Previous studies in the UIB concerning the relationship between runoff and climatic (Archer 2003), and trends in runoff (Sharif et al 2013) are confined to the latter half of the 20$^{th}$ century, with the latest records of discharge ending in 1998. Khattak et al (2011) and Siddique and Hashmi (2012), although using more recent discharge records (1967-2005) and (1995-2009), only utilise gauging stations situated along the mainstem of the Indus.

This study aims to provide a contemporary outlook on riverflows from gauging stations of glacierised tributary basins which provide significant proportions to runoff downstream with long term records extending into the 21$^{st}$ century. Since river discharge reflects climatic variability, annual totals of discharge from rivers draining basins of the Karakoram and western Himalaya are used to indirectly establish the current health of glaciers in glacierised basins.

From records of discharge the study aims to answer the following questions:

1) What are the current patterns of annual riverflows in upstream tributaries, and how are these riverflows affected downstream at large dams?

2) How do patterns of runoff at large dams change from west to east?

3) How variable are annual riverflows from year-to-year, and how does year-to-year variability of runoff change regionally?

4) Do riverflows from glacierised basins in the Karakoram indirectly support the Karakoram anomaly?

Previous studies have suggested reduced riverflows from highly glacierised Karakoram basins are attributed to reduced icemelt from declining summer air temperatures (Sharif et al 2013), thus agreeing with the Karakoram anomaly. However, (Mukhopadhyay and Khan 2014a) expressed difficulties using riverflows as indicators of glacial mass balance, suggesting a stable mass balance of Karakoram glaciers despite increasing riverflows from the glacierised Shigar basin.

From meteorological parameters similar questions were posed during the investigation:

1) What are the current patterns of seasonal precipitation totals, and how do these patterns and total quantities change from west to east and from north to south?

2) Is there any relationship between climatic variables in the UIB and western Himalaya?

3) How variable are seasonal precipitation totals from year-to-year, and how does variability change from west to east?

4) Are summer temperatures continuing to decline in the central Karakoram?

## 3. Study area

The Himalayan arc extends between 28°N-38°N and 70°E-95°E. However, this study is confined to basins in the western Himalaya and Karakoram extending between 30°N-38°N and 70°E-80°E. From west to east these basin include the Indus, Jhelum, Chenab and Sutlej. Under the Indus Water Treaty of 1960 waters of the Indus, Jhelum and Chenab were allocated to Pakistan, whilst the waters of the Sutlej were given to India, along with the Ravi and Beas. Locations of basins used in this study can be seen in figures 3-1 to 3-5. The glacierisation and altitude between each catchment varies between basins, and their basin characteristics can be seen in table 3-1. The UIB had several sub-catchments, including the Shyok, Hunza, Gilgit and Astore. Each of these basins has its own gauging station off the mainstem Indus with length of records in excess of 30-40 years, and all have considerable ice cover (Table 3-1). Gauging stations on the mainstem Indus were acquired at Besham and Tarbela dam.

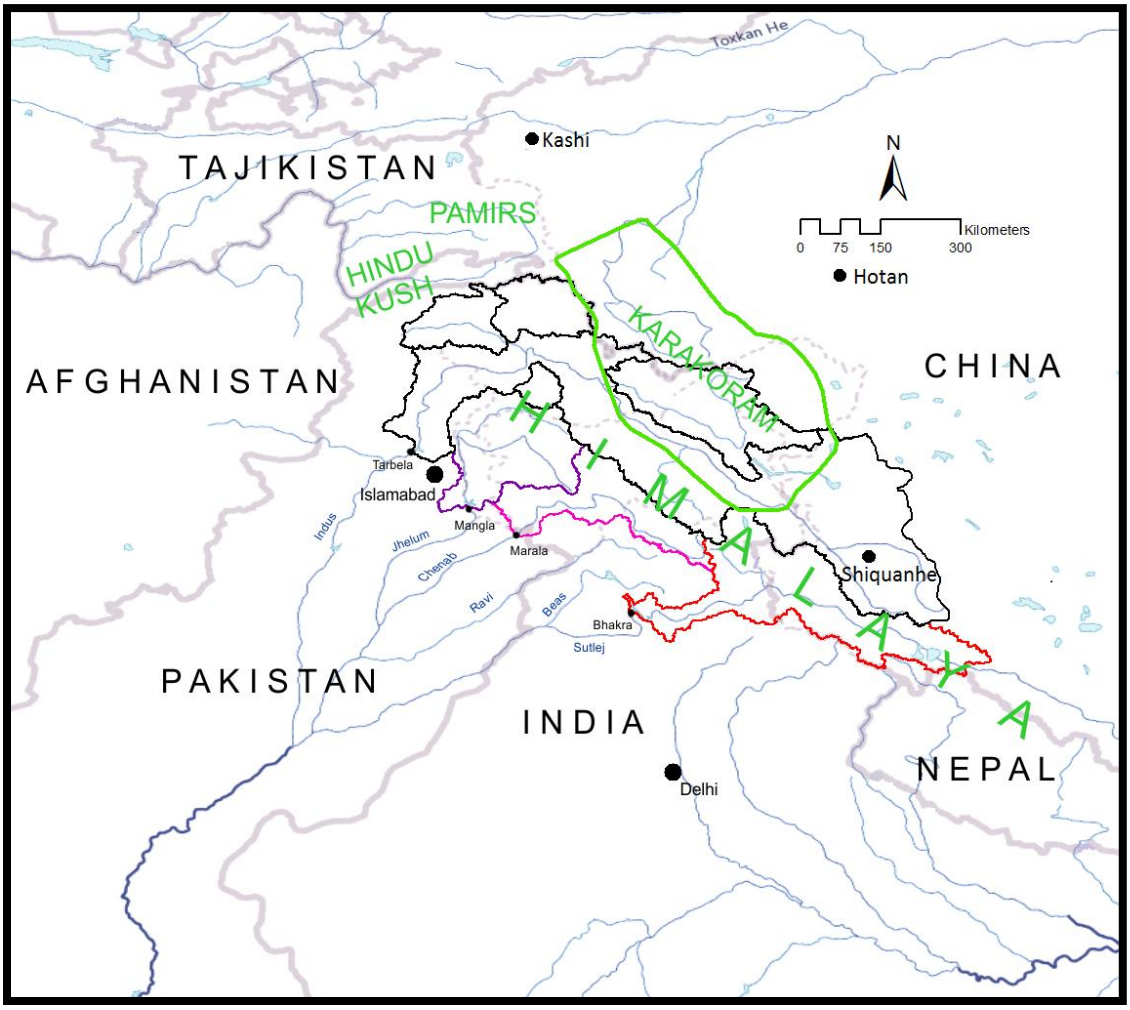


**Figure 3-1: Location of study area. Watershed outlines are indicated; UIB (black), Jhelum (purple), Chenab (pink), and Sutlej (red). Mountain ranges are indicated by green.**

Several gauging stations along the profile of the Sutlej River were obtained. From north to south these gauging stations include; Khab, Rampur, Luhri, Suni, and Bhakra Dam. For the Jhelum and Chenab basin only one gauging station was obtained for each.

Records on the Jhelum were obtained at Mangla dam with records ending in 2001, while records on the Chenab were obtained at Marala headwork, records extending 1948-2006.

## 3.1. Upper Indus Basin

The UIB can be defined by the area upstream of Tarbela dam, one of three mega-dams in northern Pakistan which regulate streamflow for irrigation on the Punjabi and Sindhi plains. A catchment area of 172,159km$^2$, the UIB, is around 8.8% glacierised with a surface area of perennial snow and ice exceeding 15,000km$^2$ (figure 3-2) (table 3-1). Rising on the Tibetan Plateau above Lake Mansarovar at 5500m above sea level the Indus is fed by small glaciers in the district of Jammu and Kashmir before entering Pakistan. The length of the mainstem Indus River is around 972km. Its major tributaries include the glacierised Shyok, Hunza, and Shigar, which have runoff peaks in August and July, and the less glacierised Gilgit and Astore with runoff maximum in July. Elevation of the UIB ranges from a maximum of 8611m a.s.l (K2) to a minimum of 336m a.s.l at Tarbela. Other major mountains include Nanga Parbat in the Astore basin (8126m a.s.l), Gasherbrum I (8080m a.s.l), Broad Peak (8051m a.s.l), and Gasherbrum II (8035m a.s.l) situated in the Baltoro glacier basin, Shigar catchment. From Tarbela Dam 35.1% of the total basin area lies above 5000m with an average elevation of 4494m a.s.l.

Over 11,000 glaciers dominate the UIB (ICIMOD 2012) but the 15 largest glaciers in the Karakoram comprise around half of the glacier area (Yao 2007 in Hewitt 2011). Some of the Largest glaciers by area include Siachen (ice area: 1181km$^2$), Baltoro (756.3km$^2$), Hispar (621.6km2) and Biafo (626.8km$^2$) (Akhtar 2008a). Surge type behaviour is common among Karakoram glaciers, and debris covers around 32% of glaciers (Hewitt 2011). Three basic types of glaciers can be recognised in the Karakoram according to Hewitt (2011):

1) The ‘Turkistan’ which has steeply incised valleys and is predominately fed by avalanched snow and ice. There is no typical accumulation zone
2) The ‘Mustagh’ also more or less entirely avalanche driven but with ice tributaries in the accumulation zone. Firn zone is sometimes identifiable
3) The ‘Alpine’: mainly driven by precipitation falling as snow in the accumulation area

Avalanching snow is the main contributor to mass, accounting for ~67% of Karakoram ice (in Hewitt 2005). 52% of total annual precipitation for the entire UIB occurs during winter months November to May, but annual precipitation recorded at valley level is typically less than ~200mm $yr^{-1}$. Precipitation is observed to increase with height in the UIB (e.g. Winiger et al 2005), and downstream as the lower elevations are unprotected by the Himalayas.

The mountainous region of the UIB remains 30-40% snow covered during the summer (Tahir 2011), but as the snowline ascends to an average elevation of 4109m a.s.l (Immerzeel et al 2009), 7000-8000$km^2$ of ice is exposed to melting (Yu et al 2013). It is this component of ice which provides 21% of total annual riverflow, along with 49% from snowmelt, at the outlet of the UIB at Tarbela reservoir (Mukhopadhyay and Khan 2014b).

Flow accumulation at Tarbela dam is primarily utilised for irrigation of agricultural land. Since the completion of Tarbela in 1977, the total irrigated area and cultivated land has increased by 12% and 35%, respectively. Tarbela represents 22% of Pakistan's installed power capacity, and it is estimated that around 40% of the population may be presently benefiting from waters supplemented by Tarbela (Siegmann and Shezad 2005). Currently, Pakistan has utilised 11% of its hydropower potential and with growing demands numerous projects are currently under feasibility in the UIB, including Thakot (4000MW) 2km from Besham, Bunji (7100MW), and Diamer Basha mega-dam (4500MW) 315km upstream of Tarbela, which is estimated to store around 15% of total annual riverflow from upstream tributaries of the Indus.

## 3.2. Jhelum basin

The catchment of Jhelum forms a natural boundary between the apex of the UIB and flows through the state of Azad Jammu and Kashmir. From Mangla dam the catchment area extends 33468$km^2$ and has a glacierisation of 0.7% (table 3-1). Headwaters of the Jhelum arise from the springs of Verinag on the north western stretch of the Pir Panjal range and from seasonal melting of 733 glaciers (ICIMOD 2012) from its northern tributaries (figure 3-3). Its largest tributary, the Neelum, ~250km in length, drains from east to west passing through Gurez valley and Dudhnial before joining the Jhelum near Muzaffarabad. The Kunhar, its second largest tributary (~168km long), drains the north-western side of the catchment passing through Naran, Kaghan and Balakot before

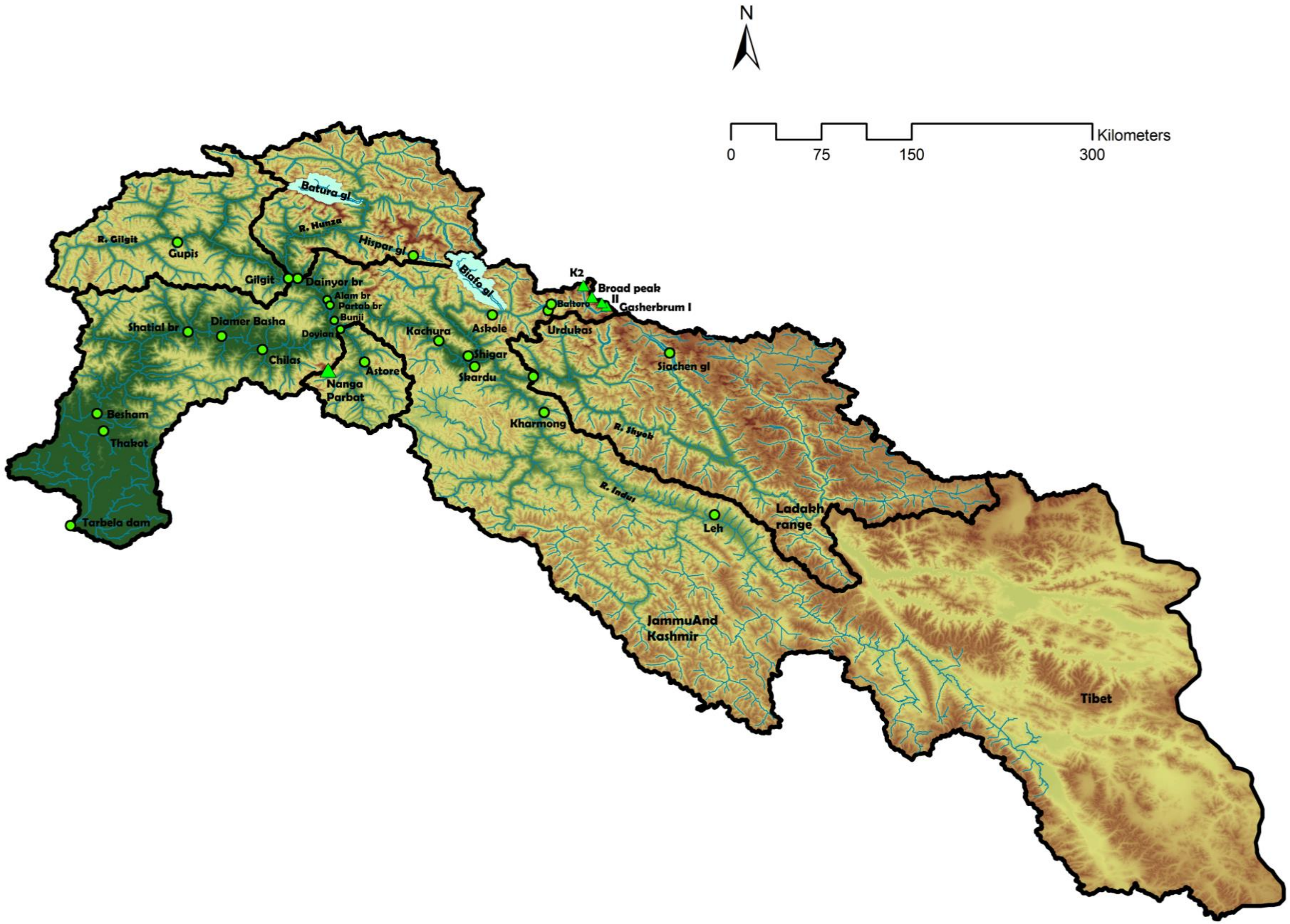


**Figure 3-2: Shuttle Radar Topography Mission (SRTM) digital elevation model (2003) of the UIB. Place names used in this study, and discussed in the text are also indicated.**

**Table 3-1: Characteristics of drainage basins in the Karakoram and Himalaya. [a] Khan et al (2014), [b] Yu et al (2013), [c] ICIMOD (2012).**

| River | Station | Period of record | Basin area ($km^2$) | Ice area ($km^2$) | Basin Glacierisation (%) | Mean runoff (m) | Elevation min (m) | Elevation max (m) | Average elevation (m) | Data source |
|---|---|---|---|---|---|---|---|---|---|---|
| **Upper Indus basin** | | | | | | | | | | |
| Shyok | Yogo | 1973 - 2005 | 33,117 | 6221[b] | 18.8 | 0.341 | 2424 | 7803 | 5093 | WAPDA |
| Hunza | Dainyor | 1966 - 2004 | 13,773 | 4339[b] | 31.5 | 0.756 | 1420 | 7809 | 4516 | WAPDA |
| Gilgit | Gilgit | 1960 - 2005 | 12,681 | 994[b] | 7.8 | 0.719 | 1454 | 7104 | 4057 | WAPDA |
| Astore | Doyian | 1974 - 2005 | 3,900 | 450[b] | 11.5 | 1.104 | 1576 | 8069 | 3995 | WAPDA |
| Indus | Besham | 1969 - 1997 | 164,853[a] | 14889[b] | 9 | 0.466 | 569 | 8566 | 4622 | WAPDA |
| Indus | Tarbela | 1962 - 2009 | 172,159[a] | 15061.7[c] | 8.8 | 0.471 | 336 | 8611 | 4494 | World Bank |
| **Jhelum Basin** | | | | | | | | | | |
| Jhelum | Mangla | 1923 - 2000 | 33,468 | 222.8[c] | 0.7 | 0.847 | 300 | 6285 | 2365 | WAPDA |
| **Chenab basin** | | | | | | | | | | |
| Chenab | Marala | 1948 - 2006 | 26,298 | 2708[b] | 10.3 | 1.238 | 228 | 7103 | 3139 | WAPDA |
| **Sutlej Basin** | | | | | | | | | | |
| Sutlej | Khab | 1972 - 2002 | 41,278 | | | 0.166 | 2503 | 7010 | 4868 | DHI |
| Sutlej | Nathpa | 1990 - 2005 | 45,907 | | | 0.285 | 1018 | 7010 | 4816 | DHI |
| Sutlej | Rampur | 1964 - 2005 | 46,857 | | | 0.237 | 434 | 7010 | 4775 | DHI |
| Sutlej | Luhri | 1972 - 2002 | 47,807 | | | 0.251 | 434 | 7010 | 4726 | DHI |
| Sutlej | Suni | 1972 - 2002 | 48,857 | | | 0.261 | 434 | 7010 | 4664 | DHI |
| Sutlej | Bhakra | 1920 - 2004 | 52,224 | 1315[c] | 2.52 | 0.297 | 337 | 7010 | 4431 | BBMB |

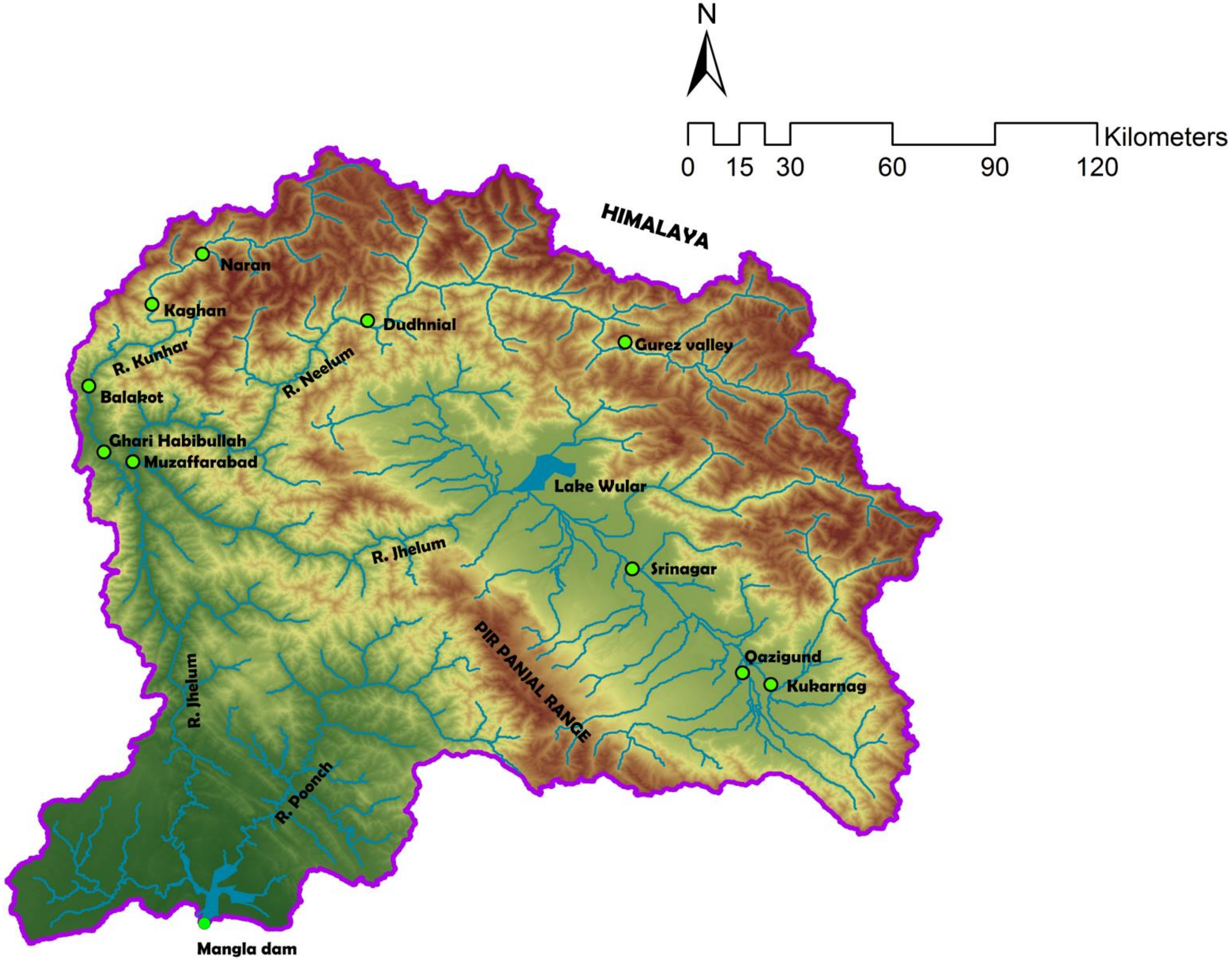


**Figure 3-3: SRTM digital elevation model (2003) of the Jhelum catchment. Place names used in the study, and discussed in the text are also indicated.**

meeting the confluence of the Jhelum 12.8km downstream of Muzaffarabad. Part of the Jhelum flows west through the alluvial plains of Kashmir valley passing through Srinagar and Lake Wular before joining the western mainstem at Muzaffarabad. It smallest tributary, the Poonch, drains the southeast Jhelum, derived from waters draining the southern foothills of the Pir Panjal before discharging into Mangla reservoir.

Emerging on the southern foothills of the Himalaya 0.06% of the basin area above Mangla lies at an elevation greater than 5000m a.s.l. The elevation of the basin ranges from 300m a.s.l to 6285m a.s.l, with an average altitude of 2365m a.s.l. As a result of the south lying catchment average annual total precipitation is around 804mm. Around 264mm (32.8% of total annual precipitation) is derived during the summer monsoon period, June to September, largely penetrating the lowland areas. In contrast to the UIB 24.9% of total annual discharge is acquired from snowmelt, but its largest proportion of annual discharge (65.1%) is derived from rainfall between the months May to October (Bookhagen and Burbank 2010). High evaporation within the catchment accounts for -17% of total annual discharge (Bookhagen and Burbank 2010) but with an early transit of spring snowmelt monthly discharge peaks in May for parts of the Jhelum and June for the Neelum and lower Kunhar. Discharge at Mangla has its highest proportion of total annual runoff in July, followed by June, whilst the Poonch, which is largely influenced by monsoon climax' in August and September (Archer and Fowler 2008).

## 3.3. Chenab basin

Marala barrage, situated on the River Chenab is located 90km southeast of Mangla dam at latitude 32° 74’N and longitude 74° 30’E. At present, 10.3% of its 26,298km$^2$ catchment area above Marala is covered with perennial snow and ice. The basin extends from 228m a.s.l to 7103m a.s.l with an average elevation of 3139m a.s.l. Over 2000 glaciers are situated in the Chenab basin having elevations in range of 3001-7103m a.s.l (ICIMOD 2012). 11.6% of the basin area above Marala lies at an elevation of 5000m or greater, with higher elevations being confined to the north of its basin. Its highest mountains are situated in the Lahaul and Spiti, Himachal Pradesh, with many of its highest peaks extending over 6000m a.s.l. Some of its highest peaks include; Mulkila (6517m a.s.l), Papsura (6451m a.s.l), Menthosa (6443m a.s.l) (Figure3-4). Headwaters of the

Chenab, namely the Bhaga and Chandra, emerge from the Baralacha Pass, Lahaul, Himachal Pradesh, India, from seasonal snows and small glaciers in the eastern Chenab. The two rivers flow in an easterly direction draining the Himachal Pradesh and Himalaya before conjoining to form the Chenab at Tandi in the state of Jammu and Kashmir. The Chenab flows its course northwest draining smaller tributaries before turning south for 27km after joining the Marusudar at Bandarkoot near Kishtwar. Thereafter the Chenab flows west 120km through Baglihar dam in the district of Doda before again turning south at Salal Dam. 35km downstream of Salal, the Chenab passes through Akhnoor gauging station, before reaching Marala 38km downstream near the industrial city of Sialkot, Pakistan. Average annual total precipitation in the Chenab catchment is about 720mm but varies considerably. Northern parts of the basin receive snowfall during the winter along the Himalayan Mountains, whilst the middle and lower reaches receive a mixture of snow and rain depending on meteorological conditions (Singh et al 1997). South facing slopes are generally wetter but also receive higher radiation, thus greater melt rates (Singh et al 1997). Winter precipitation from November to May and summer precipitation from June to September as an average for the entire basin account for nearly equal quantities, 52.6% and 43.3%, respectively. Around 70% of the basin is covered with seasonal snow at the end of winter, but reduces to ~24% at the end of the ablation season in September and October (Singh et al 1997). Almost half of total annual discharge is derived from snow and glacier ice melt, estimates varying between 43.4% (Bookhagen and Burbank 2010) and 49% (Singh et al 1997). Compared with its less glacierised neighbouring catchment, the Jhelum, evapotranspiration is considerably less accounting for -7.6% of total annual discharge (Bookhagen and Burbank 2010). Riverflow at Akhnoor peaks in July, followed by a secondary maximum in August, from a combination of monsoon rains and glacier melting, post-monsoon (Singh et al 1997).

## 3.4. Sutlej basin

Rising From Lakes Mansarovar and Rakastal in Tibet the Sutlej flows west for about 295km before entering India at Shipkila in the state of Himachal Pradesh. The Sutlej is joined by its main tributary, the Spiti, at Khab upon entering India and takes a south-westerly course where it is fed by smaller tributaries such as the Ropa, Kashang and Baspa, above the Nathpa Jhakri Dam. From Nathpa the Sutlej flows about 43km southwest before passing Rampur, and a further 31km

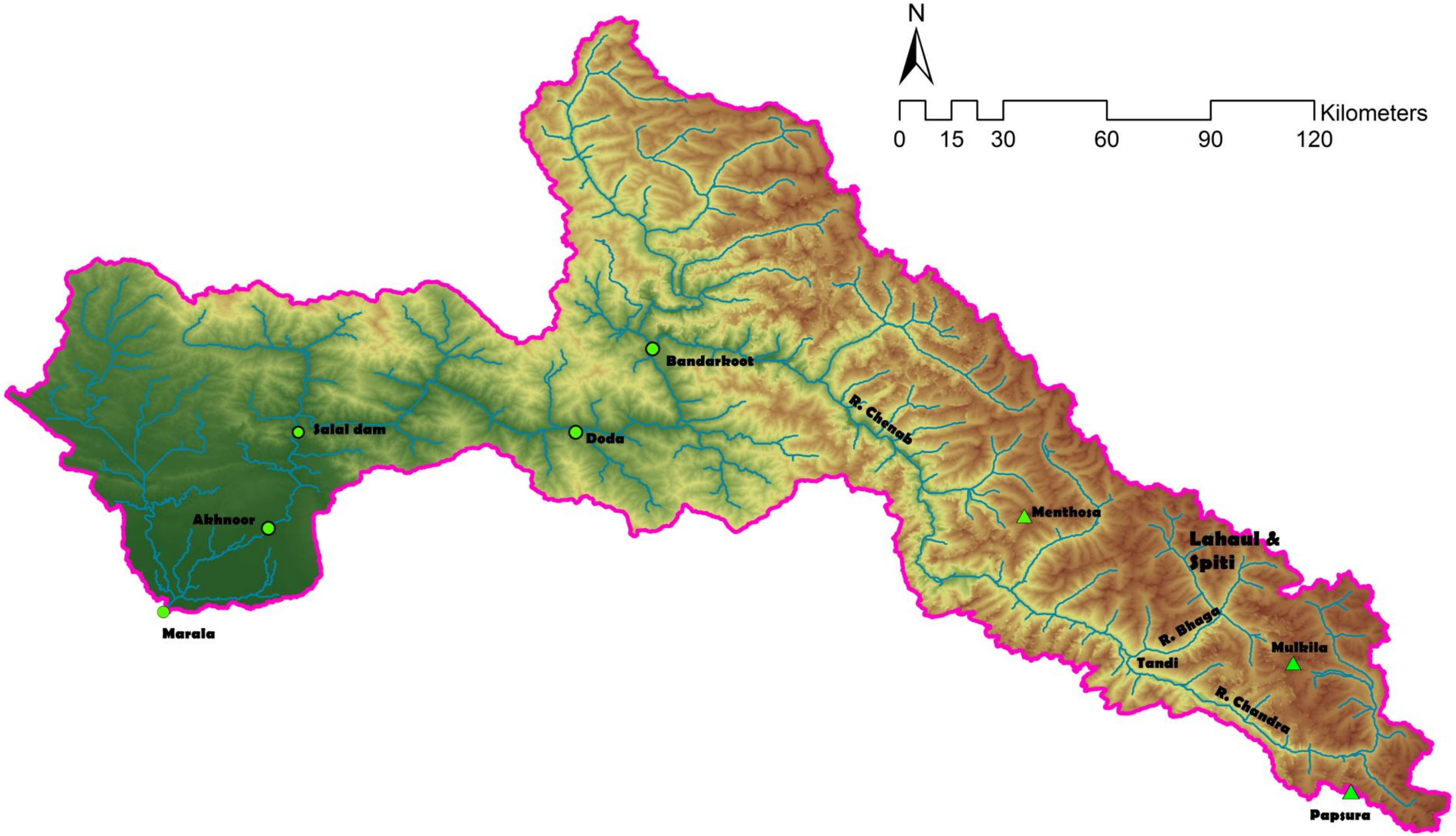


**Figure 3-4: SRTM digital elevation model (2003) of the Chenab catchment. Place names used in the study, and discussed in the text are also indicated.**

before reaching Luhri. 47km pass from Luhri to Suni, and 128km from Suni to Bhakra Dam (figure 3-5).

Currently 2.52% of the catchment above Bhakra Dam is glacierised, with a catchment area extending 52224km$^2$. 2108 glaciers cover the Sutlej basin with an area of 1315km$^2$ being covered with permanent snow and ice (table 3-1). Elevation of the basin area above Bhakra extends 337m a.s.l to 7010m a.s.l with a mean of 4431m a.s.l. Highest elevations are confined to the north in the Himachal Pradesh where the lowest glacier termini reach around 3600m a.s.l (ICIMOD 2012). From Khab 38% of the basin lies above 5000m a.s.l, but decreases to 30.7% from Bhakra as elevation decreases with distance from the Himalaya. On average 65% of the total basin area remains snow covered at the end of winter (Singh and Jain 2002), also confined to north of the basin at higher elevations. Snow cover reduces to around 20.3% in September (Singh and Jain 2002), but the lower elevations (below Rampur) remain snow free throughout the entire year.

On average the Sutlej basin is generally arid (average precipitation 611mm yr$^{-1}$ in the period 1950-2000) as ~57% of the basin area above Bhakra dam lies in arid Tibet (calculated from Singh and Jain 2002). Annual totals of precipitation for the basin area above Khab, which lies almost entirely in Tibet, is around 486mm yr$^{-1}$. 22,305km$^2$ of the basin area above Bhakra lies in India (Singh and Jain 2002) which receives higher quantities of precipitation, annual precipitation totals between Rampur and Bhakra ranging between 800mm and 1500mm yr$^{-1}$ (Singh and Jain 2002). Seasonal totals of precipitation over the Sutlej are slightly biased toward the monsoon which accounts for 57.6% of total annual precipitation, while winter precipitation from November to May accounts for 37.8%. As a result 41% of total annual discharge at Bhakra is derived from rainfall, but the largest proportion (59%) is derived from snow and ice melt with peak riverflow in July, followed by August (Singh and Jain 2002).

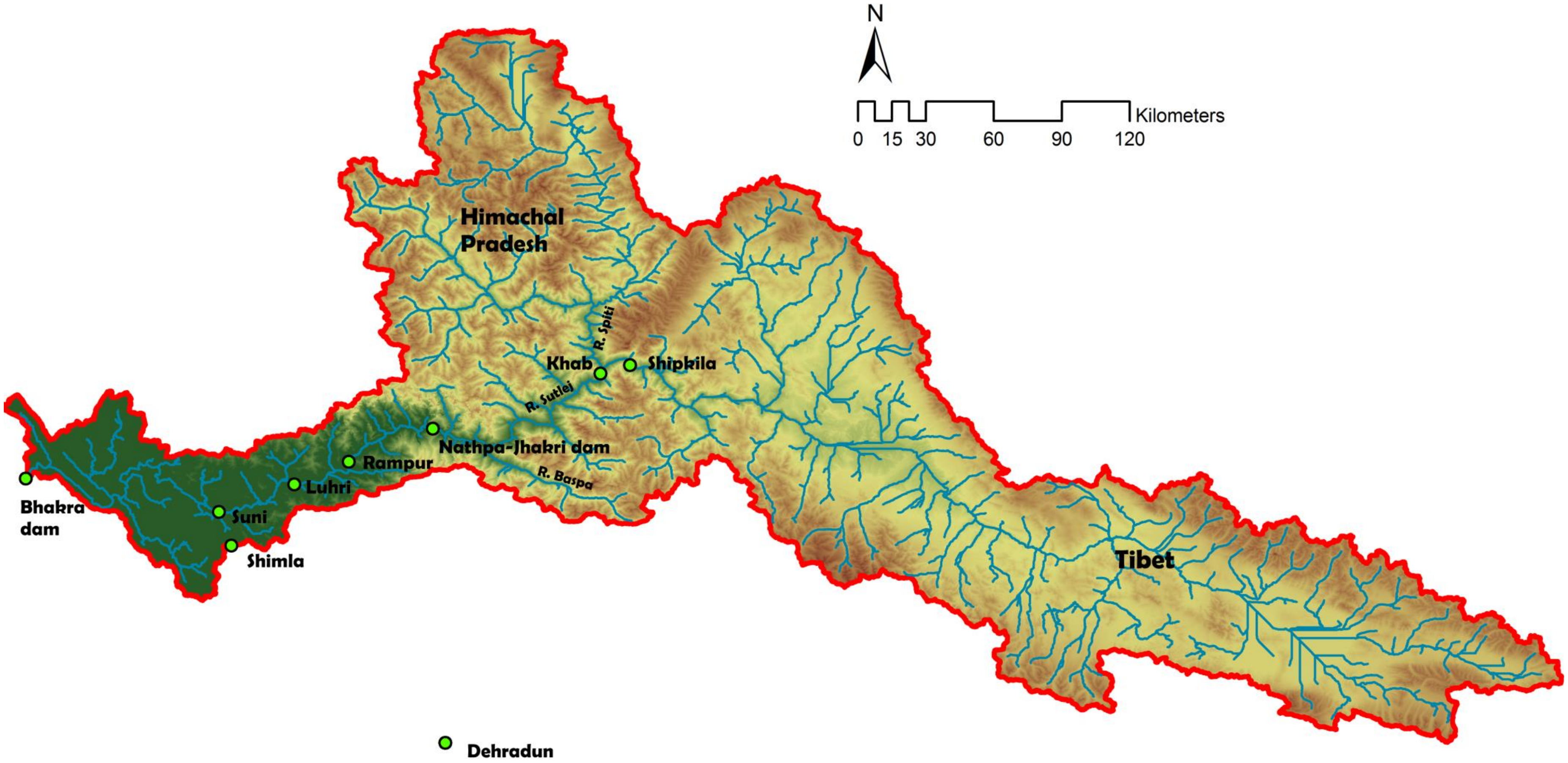


**Figure 3-5: SRTM digital elevation model (2003) of the Sutlej catchment. Place names used in the study, and discussed in the text are also indicated.**

# 4. Literature review

## 4.1. Sources of runoff from glacierised catchments

Total yield of discharge gauged downstream of a partly glacierised basin is made up of two distinct components (e.g, Chen and Ohmura 1990):

1) Runoff from the ice free area
2) Runoff from the ice area

The ice free area contains no perennial snow and ice and usually consists of lower elevations of the glacier basin. Water derived from the ice free area can be divided into three components (Collins 1979):

1) Snow melt
2) Rainfall
3) Groundwater

The glacier area, by contrast, is confined to higher elevations and has a more stable snow cover (Chen and Ohmura 1990). Runoff derived from the glacier can be divided into several components (Collins 1979) including:

1) Snowmelt on the glacier
2) Surface ice ablation
3) Subglacial melting
4) Precipitation runoff
5) Groundwater
6) Internal water from ice deformation and frictional melting

The most important sources of runoff which yield the greatest quantities of water from the glacier area are surface ice ablation and snowmelt on the glacier.

The transition of runoff components into actual streamflow varies both seasonally and annually depending on meteorological conditions within the basin. Runoff from snowmelt mainly occurs during spring and early summer when a rise in radiation allows for conditions to favour melting of the winter snowpack. After the winter snowpack has melted, surface ice melting contributes significantly to runoff from mid to late summer.

## 4.2. Climate and runoff relationship in glacierised catchments

It is well established that the ice area of a catchment is positively related to air temperature and negatively related to precipitation (Collins 1987, 1989). Likewise, the ice free area is positively related to precipitation and negatively related to air temperature. Runoff from the ice free area is derived from and directly related to, but always less than, total precipitation. However, runoff from the ice area can be less than, equal to, or greater than total precipitation (e.g, Collins 1989). As a function of the total basin, the relationship between climate and runoff is moderated by opposing responses of runoff from the ice free area and ice area to the same meteorological inputs. The percentage glacierisation of basin (Ice area/ total basin area) can determine the relationship between climate and runoff.

Glacierisation of a catchment can be characterised into four general categories:

1) Ice free: basin contains no glacier(s)
2) Low glacierisation: basin has a limited ice cover/is near ice free
3) Intermediate/moderate glacierisation: basin has near equal proportions of ice area and ice free area
4) Highly glacierised: basin area is largely dominated by ice and perennial snow. A small proportion of the basin remains ice free

In an ice free basin, total annual yield of runoff is always less than total precipitation, but year-to-year variations of runoff are related to, and reflect seasonal variations of precipitation (Figure 4-1) (e.g, Collins 2007). Winter precipitation provides bulk of the riverflow predominantly from snowmelt and sustains riverflows throughout spring and early summer (Collins 2006, Fountain and Tangborn 1985). Runoff in an ice free basin is reduced during the summer season and in warm dry years as the seasonal snowpack is limiting and there is no glacier to provide additional runoff, although Himalayan basins which receive precipitation from the monsoon have enhanced riverflows from June to September (figure 4-2) (e.g, Archer 2008, Collins 2013).

As basin glacierisation increases, the association between runoff and air temperature increases as the surface area of ice available for melting is greater (figure 4-1). Volume of annual discharge generally becomes greater, as does the proportion of summer discharge to total annual discharge (e.g, Collins 2006, Escher-Vetter and Reinwarth 1994). Maximum monthly discharge is delayed further into summer as the glacier area

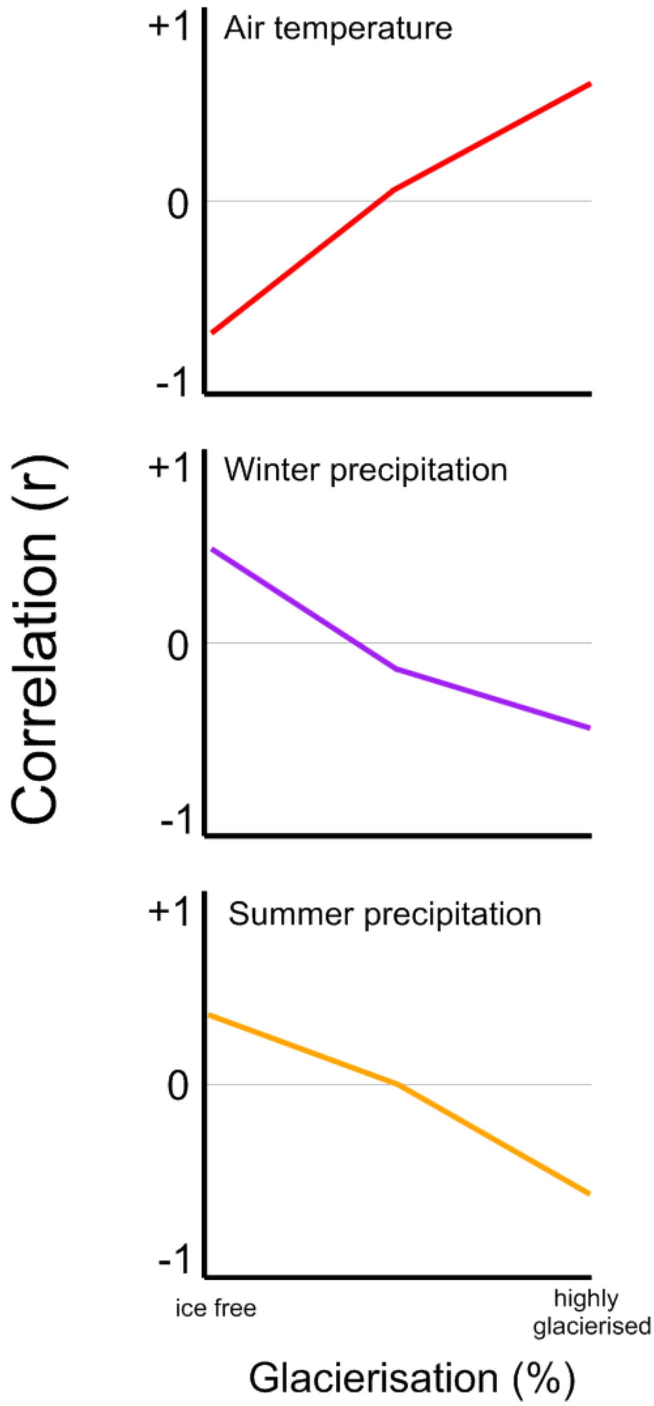


**Figure 4-1: Schematic diagram showing the correlation between discharge and climatic variables against the percentage glacierisation of a catchment.**

increases, until the internal drainage system of the glacier becomes favourable to transit stored meltwater (Figure 4-2) (Jansson et al 2003). From glacierised Alpine basins in the Swiss Alps Collins (2006) noted, with increasing glacierisation, 85.1%, 88.2%, 92.3% of total annual runoff occurred during the summer months May to September, with

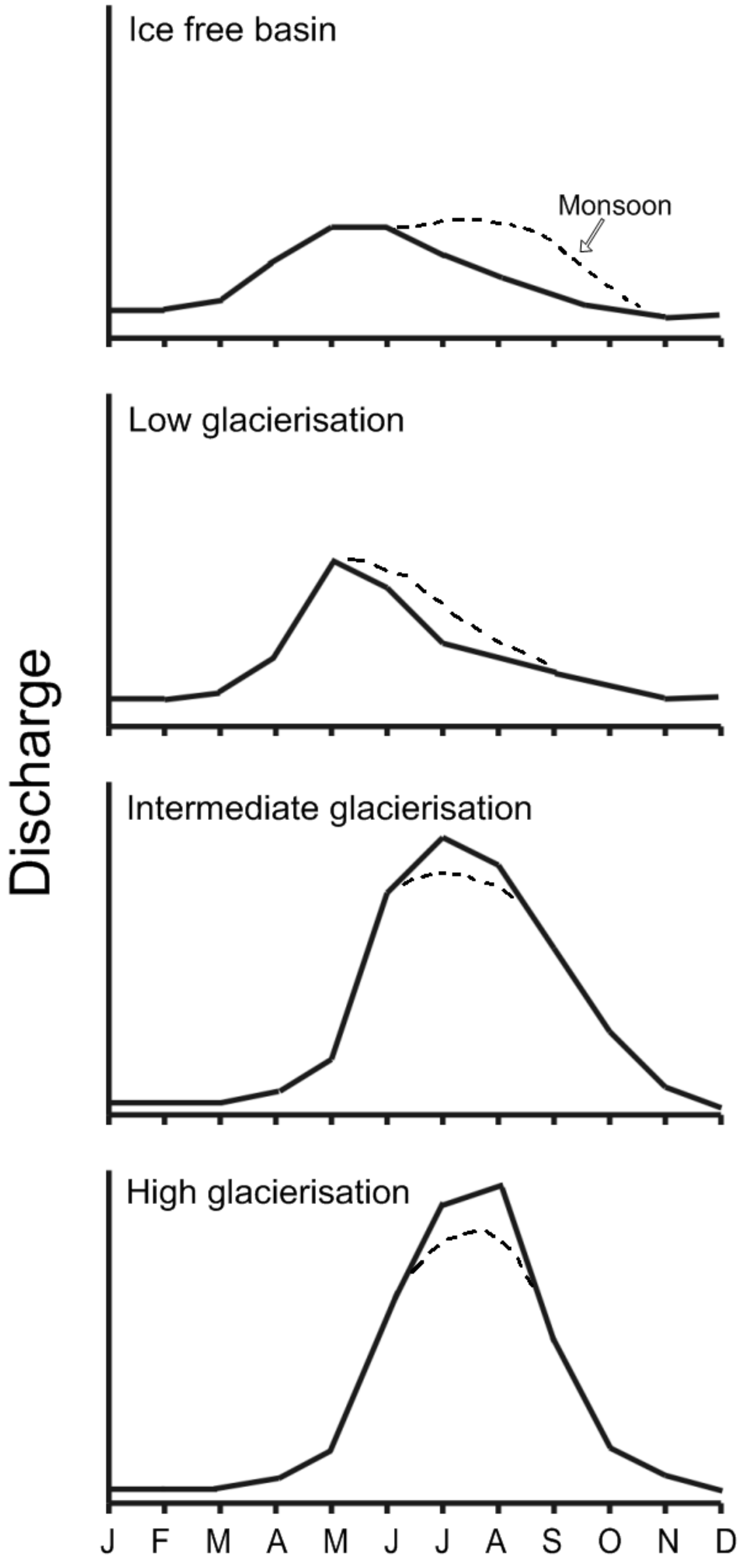


**Figure 4-2: Schematic representation of hydrographs from mountain basins with varying proportion of ice cover.**

maximum monthly discharge being delayed in July for a 36.5% and 52.2% glacierised basin, and August for a 65.9% glacierised basin.

Moderately glacierised basins have annual variations of runoff which are intermediate between annual variations of energy availability and precipitation, and are neither dominated by one climatic parameter (Figure 4-1) (e.g, Collins 1987, Collins 2007, Fountain and Tangborn 1985). Annual totals of runoff may be equal to, less than, or more than, annual totals of precipitation depending on meteorological conditions within the glacier basin which affect storage or water as glacier ice (Collins 2006). A tendency to higher (lower) glacierisation will have a stronger (weaker) relationship with air temperature and a weaker (stronger) relationship with precipitation. Highly glacierised basins have patterns of annual riverflow which strongly reflect annual variations of energy availability for melting, and high runoff totals may be superimposed under a period of sustained climatic warming (Collins 2007, 2008). In highly glacierised basins maximum monthly riverflow is delayed until August when the seasonal snow cover is at its minimum, the transient snowline is at its highest elevation, and the drainage network allows from rapid transmission of meltwater from storage (Jansson et al 2003). Winter precipitation provides additional runoff from melting of snow over both the ice and ice free area, but particularly wet winters may reduce total runoff as a thicker layer of snow delays the rise of the transient snowline and reduces the area and duration of ice exposed to melting during summer, such that the relationship between winter precipitation and runoff may become negative (figure 4-1). Summer precipitation has similar effects, precipitation events reduce incoming solar radiation which limits surface melting of the glacier, and precipitation falling as snow has a higher albedo than underlying ice which may produce recession flows (Collins 1982, Rothlisberger and Lang 1987). Summer precipitation events can cause outburst floods, particularly if the basin is snow free as rainfall on ice or moraines will runoff rapidly (Collins 1998). From year-to-year warmer summers generally yield higher runoff totals in glacierised basins compared with winters that have higher snowfall as the melt per unit of ice area is greater than that of snow.

## 4.3. Variability of annual runoff 'The compensation effect'

The variability of annual runoff, calculated using coefficient of variation (CV), is reduced in basins with small proportions of glacier cover compared with ice free basins as icemelt moderates runoff in drier years when precipitation totals are low (e.g, Collins 1990, Tvede 1982, Krimmel and Tangborn 1974, Fountain and Tangborn 1985). Variability of annual and summer runoff reduces with increasing glacier cover to a minimum around 20-40% glacierisation, but then increases in basins which are dominated by ice cover, creating a U-shaped relationship between percentage glacierisation and CV (figure 4-3) (Chen and Ohmura 1990, Collins 1990, 2006, 2007, Fountain and Tangorn 1985).

Intermediately glacierised basins tend to reduce variability of runoff to a minimum by dampening extreme climatic events which produce opposing runoff responses from ice and ice free areas. In warm dry summers following winters with little snowfall, runoff from the ice free area is reduced, but with a thinner winter snowpack the transient snowline will rise sooner, and higher, exposing a larger surface area of ice to melting. Runoff from the ice area will be greater and compensate for reduced runoff from the ice free area. Conversely, in cooler-wetter years following high winter snowfall the timing of the transient snowline will be offset and will rise later, encompassing a smaller surface area of ice to melting during summer. Avalanching snow from surrounding slopes will be greater and provide an even greater protective blanket of snow over the glacier surface (Krimmel and Tangborn 1974). The release of melt water from the glacier will be reduced, but runoff from snowfall and liquid precipitation will compensate for the ice melt deficit irrespective of spring and summer temperatures which affect only the timing not total quantity, of runoff (Ferguson, 1985). From a study on South Cascade Glacier, USA, Krimmel and Tangborn (1974) demonstrated runoff was approximately the same in two years with contrasting snow cover, while other parameters such as air temperature, cloud cover, and precipitation were similar.

Variability of annual runoff increases in ice free and highly glacierised basins as the compensation effect from opposing runoff is reduced. A Study in Northwest China (Zuming 1982) demonstrated annual runoff from the ice free Xida and Gulang basin varied by 41% and 20.9% in drought years, while in a wetter year runoff varied by 39.4% and 53%. Highly glacierised basins are dominated by ice melt, providing plentiful runoff in warm years. In cooler years, however, icemelt runoff is subdued, but the ice free area over which precipitation compensates for runoff is limited. Ice free basins generally have higher variability than highly glacierised basins since intra-annual variations of precipitation are higher than energy fluxes (e.g, Collins and Taylor 1990).

Variability of runoff can also change considerably between regions irrespective of basin glacierisation as precipitation is spatially heterogeneous (Moore 1992). A study in Norway demonstrated higher annual variability of runoff in the drier south-east where convective storms were common, whilst lower CV of annual runoff was common in the wetter, western region of Norway where frontal precipitation dominates (Tollan 1972).

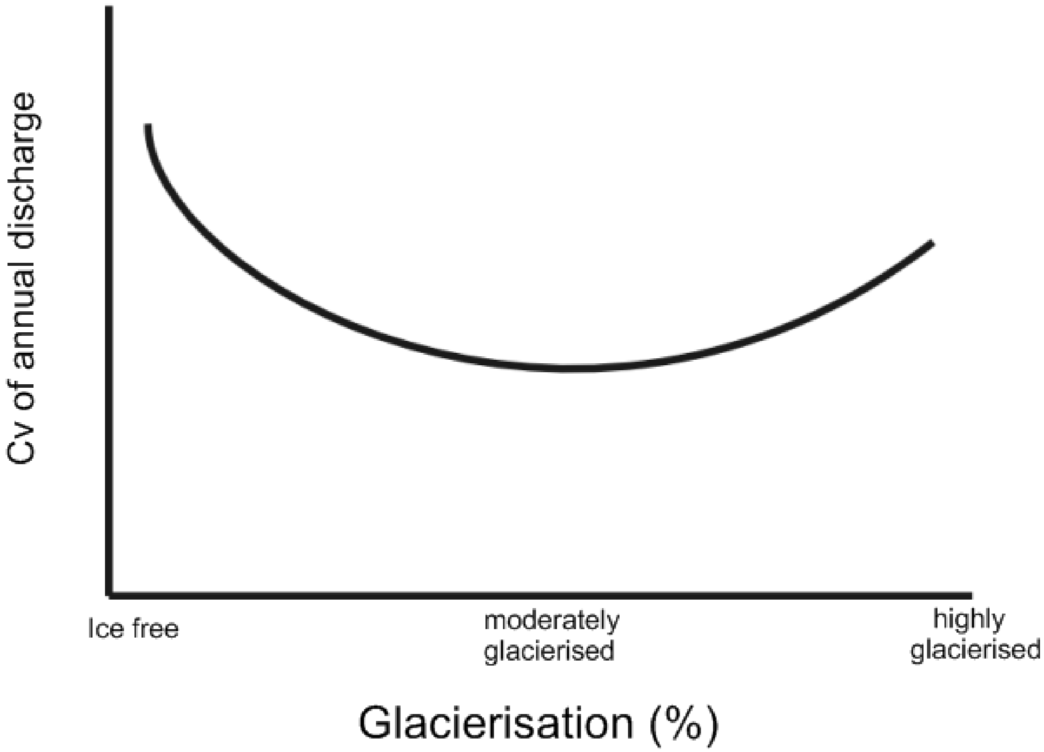


**Figure 4-3: Schematic diagram showing the relationship between CV of annual discharge and percentage glacierisation of a drainage basin.**

## 4.4. Climatic regime in the Himalayas

During winter and spring the Karakoram-Himalaya is frequently under the influence of the Tibetan anticyclone and conditions are usually cool and dry. However, low pressure–westerly disturbances originating from the Mediterranean and Atlantic often track eastwards and transfer moist air to the Hindu Kush, Karakoram and western Himalaya, causing heavy precipitation from October to March (Ridley et al 2013). In the Karakoram around 67% of precipitation is derived during winter and is intensified with elevation and large topographic features (Hewitt 1989); with monthly maximum precipitation and snow cover occurring in March (Akhtar et al 2008b, Immerzeel et al 2009). Winter precipitation decreases rapidly eastward, where precipitation from November to May at Shimla accounts for only 25% of total annual precipitation (Collins et al 2013).

Connections between the North Atlantic Oscillation (NAO), determined by differences in normalised sea level pressure over the Azores and Iceland, and winter precipitation in the UIB has been investigated previously by Archer and Fowler (2004) and Minora et al

(2013). From 1960-1999 Archer and Fowler (2004) found significant positive correlation between monthly precipitation and monthly NAO, particularly for northern climate stations, Astore, Bunji and Skardu. Highest correlations were observed between a November to January average NAO index and December to February precipitation at climate stations Astore and Bunji, but generally decreased in stations located south of the Himalaya. Positive correlation between NAO and winter precipitation in the UIB has also been observed with more recent data (1980-2009) but correlations were weaker in value, strongest correlations being between NAO and precipitation between October and December (r= 0.17) (Minora et al 2013).

During the summer months a strong pressure gradient resulting from the thermal field of the Tibetan Plateau modifies the westerly airflow into a south-easterly wind system (Bookhagen and Burbank 2010). Monsoon vortices are formed along the Bay of Bengal and move north and northwest, resulting in intensified rain between the months June to September. The monsoon decreases from east to west (figure 4-4), most notably along the southern foothills, where a 6-fold east-west gradient is observed (Bookhagen and Burbank 2010). Along the main stretch of the Himalaya (500-5000m a.s.l) rainfall amount is less but remains fairly uniform from east to west. The Sutlej and the Garhwal are identified as transitional zones, where precipitation is divided between winter westerlies and eastern monsoon (Bookhagen and Burbank 2010, Barros et al 2004). To the west 90% of the UIB lies in the rain shadow of the Greater Himalaya (Immerzeel et al 2009), monsoon incursions are weak (figure 4-4), with a slight secondary maximum between June and September (Mukhopadhyay and Khan 2014b), accounting for around 28-33% of total annual precipitation in the Karakoram (Fowler and Archer 2004, Wake 1989). However, not all summer precipitation in the Karakoram is of monsoon origin. Moisture drawn from the Arabian Sea can deliver precipitation to the Karakoram from late spring to early summer (wake 1987, 1989). Archer and Fowler (2004) also suggested westerly weather systems influenced summer precipitation in the Karakoram from Negative correlations between NAO and summer precipitation. However, Minora et al (2013) found a slight positive correlation between NAO and July to September precipitation in the UIB (r=0.18).

Precipitation is strongly influenced by elevation in the Himalayas, and is enhanced by topographical features on southward facing slopes (Bookhagen and Burbank 2006). This phenomenon particularly applies for the Karakoram, where meteorological stations at ~2000m a.s.l record annual totals around 100-150mm (e.g, Archer and Fowler 2004). Figure 4-4 shows the aridity at low elevations across the UIB, Tibet and most of the western Himalayas. From glaciological studies of Biafo glacier, Wake (1989), estimated maximum accumulation occurred at elevations 4900-5100m a.s.l. Estimates from Baltoro near K2 suggest precipitation increases to 1600mm yr$^{-1}$ at 6100m a.s.l, and may further increases to 2500mm yr$^{-1}$ at 8000m a.s.l (Decheng 1978: cited in Mayer et al 2006). Immerzeel and others (2012) estimated maximum precipitation in the Hunza basin occurred at 5500m a.s.l, and total basin precipitation was 260% higher than what was observed at valley stations. They further suggested the vertical precipitation lapse rate was 0.21 ± 0.12% m$^{-1}$, but above 5500m precipitation lapse rate reversed. Further south at Nanga Parbat precipitation is marginally influenced by the monsoon and precipitation totals reach around 2300mm yr$^{-1}$ at 5500m a.s.l (Winiger et al 2005). Along the central Himalaya maximum precipitation occurs at much lower elevations, maximum elevation being between 3000-4000m a.s.l (Bookhagen and Burbank 2006, Putkonen 2004). Despite low annual totals of precipitation observed at valley stations in the UIB, especially during summer, snow cover decreases from west to east in all seasons (Immerzeel et al 2009).

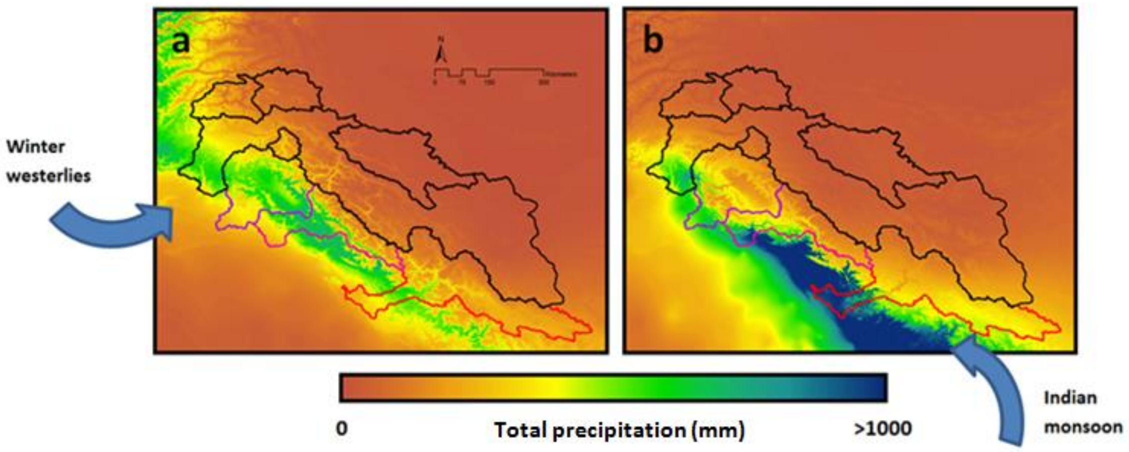


**Figure 4-4: (a) total winter precipitation (mm) from November to May (b) total summer precipitation from June to September (mm) across the Karakoram Himalaya. Data are based on monthly average gridded precipitation from valley stations between 1950 and 2000 courtesy of Worldclim.org.**

## 4.5. Precipitation trends

### 4.5.1. Winter precipitation

From climate stations in the northern UIB, a slight decline of winter precipitation from 1961-1980 was offset by a large increase since the mid-1980s (Archer and Fowler 2004). From 1961-1999, October to March precipitation significantly increased at Skardu by a rate of 16% per decade, whilst neighbouring meteorological stations Gilgit, Astore and Bunji only had slight, non-significant increases. Significant increases of January-December precipitation were noted for Gilgit and Bunji, with increases of 10% and 13% per decade (Archer and Fowler 2004). Khattak et al (2011), with slightly longer records of meteorological data (1967-2005), observed steady winter precipitation rates over a 39 year period for the Upper region of the Indus basin around Gilgit-Baltistan, with a non-significant increase of 6.25mm over 39 years. However, in the middle reaches of the Indus basin, (areas surrounding Tarbela) winter precipitation was observed to decline at a rate of -11.23mm over 39 year (1967-2005). Similar observations were made by Bocchiola and Diolaiuti (2013); totals of January to March precipitation from 1980-2009 slightly declined at Astore and Chilas, but increased at Bunji and Gilgit, with statistically significant increases at Gupis. The number of wet days for the same period increased over upper reaches of the Indus, as did cloud cover, which has been increasing since 1991 at Gupis. From a combination of satellite rainfall estimates and gridded in-situ rain gauge data Palazzi et al (2013) noted long term trends of December-April precipitation in the Hindu-Kush Karakoram had no significant trends. However, precipitation appeared to increase from the mid-1980s before declining from the mid-1990s to a minimum in the early 2000s.

In contrast to steady November-April snowfall in the Karakoram from 1991/92 – 2006/07, snowfall in the Pir Panjal range and Greater Himalaya declined by ~280 and 440cm between 1988/89 and 2007/08 (Shekhar et al 2010). With shorter records, total winter precipitation between November to April declined at Leh in the eastern Indus catchment, but increased at Srinagar in the Jhelum basin from 1901-1989 (Bhutiyani et al 2010). Decreases of winter precipitation at Srinagar between 1903 and 1982 were offset by increases from 1962-2002 (Kumar and Jain 2010). East of the Sutlej River, total winter precipitation from November to May at districts Shimla and Dehradun declined from the early 1900's to a minimum in the 1960's and late 1990's with temporary recovery in the 1980's (Collins et al 2013). Winter precipitation for the western Himalaya as a whole had a positive trend from 1866-2006 according to Bhutiyani et al (2010), however from 1988/89 to 2006/07 winter snowfall from the western Himalayas declined by 280cm (Shekhar et al 2010).

### 4.5.2. Summer precipitation

Summer precipitation from northern stations in the UIB displayed significant increases at Astore, Bunji and Skardu from 1961 to 1999, intensifying during the 1990's, with decadal increases of 13, 9 and 6mm respectively (Archer and Fowler 2004). This is supported by Khattak et al (2011) who observed a statistically significant increase of summer precipitation in the upper region of the UIB at a rate of 23.9mm over 39 years, 1967-2005. In the middle reaches of the Indus, around Tarbela, summer precipitation for the same period declined by -52.10mm over 39 years (1967-2005), although the trend was not statistically significant. From 1980-2009, northern meteorological stations displayed stationary behaviour, with some negligible increases at Bunji, Gupis and Astore, and some minor decreases and Gilgit and Chilas, none of which were significant (Bocchiola and Diolaiuti 2009). Similar observations were made by Palazzi et al (2013). With varying lengths of records, totals of June to September precipitation over the Hindu-Kush Karakoram had no specific trends. From year-to-year precipitation varied considerably, declining in the 1960's and 1980's, before intensifying in the 1990s. Following the late 1990 maxima, precipitation declined considerably in the early 2000s.

At the lower elevations of Kashmir valley, Jhelum basin, non-significant increases of summer monsoon were observed at Srinagar and Qazigund, but a decrease was observed at Kukarnag between 1962-2002 (Kumar and Jain 2010). Annual rainfall appeared to intensify in the 1990's but declined in the late 1990's to early 2000's (Kumar and Jain 2010), similar to the pattern in the Indus catchment. To the east, total July-September precipitation declined at Muzaffarabad (Bochiola and Diolaiuti 2013) and precipitation from June to September generally declined from the 1960's to the early 2000 at Shimla and Dehradun, where precipitation at Dehradun district declined by 30-40% from 1960-2000 (Collins et al 2013). Bhutiyani et al (2010) reported a significant decline in monsoon precipitation (June-September) at Shimla and for the entire western Himalaya region between 1866 and 2006.

## 4.6. Air temperature trends

### 4.6.1. Summer air temperature

Important for melting of perennial snow and ice, average June-September air temperature increases have been reported for the northwest Himalaya region between 1876 and 2006 (Bhutiyani et al 2010). Likewise, significant warming rates during the monsoon period have also been reported at Shimla between 1876 and 2006, and at Leh and Srinagar between 1901 and 1989, although not significant at Srinagar (Bhutiyani et al 2010). Collins et al (2013) reported general warming of May-September mean air temperature from the 1990s to early 2000's for central and eastern Himalayan meteorological stations, but with limited records from western Himalayan climate stations, air temperature inclined from the 1970s before ending in the 1980s.

On the contrary, mean July-August air temperature from northern climate stations in the UIB displayed cooling trends from 1961-1999, with significant decreases at Bunji, Gilgit and Astore (Fowler and Archer 2006). Cooling of average monsoon air temperature in high mountain climate stations of the Karakoram-Himalaya was also observed by Hussain et al (2005) between 1971 and 2000, but subsequently increased in the sub-mountain region along the southern slopes of the western Himalaya and Hindu Kush. From short, recent records (1995-2009), Siddique and Hashmi (2012) observed significant falling trends of positive degree days from WAPDA installed weather stations in the Hunza, Gilgit and Astore basin during the glacial melt period, July-September, whilst rising trends were observed during early summer, May-June. Significant decreases in July-August minimum temperatures were observed at Gupis between 1980 and 2009 but only slight (non-significant) decreases were observed at Bunji and Chilas (Bocchiola and Diolaiuti 2013). Mean maximum summer air temperatures from July-August displayed no specific trend at individual stations in the northern Indus basin (Bocchiola and Diolaiuti 2013), however a negative (non-significant) cooling trend was observed for the entire upper region of the Indus between 1967 and 2005 (Khattak et al 2011).

## 4.7. Riverflow trends from partially glacierised basins

### 4.7.1. Annual riverflow

Annual totals of riverflow which include disproportionate quantities of discharge derived during winter, spring, summer, and autumn have been studied in the western Himalaya by Bhutiyani et al (2008). From west to east, riverflow on the Chenab at Akhnoor and Ravi at Madhopur have increasing annual trends from the 1960s to 1990s, whilst decreasing trends have been observed on the Beas at Thakot and Sutlej at Bhakra (Bhutiyani et al 2008). Bhutiyani et al (2008) attributed trends at Ravi and Beas to changing patterns of monsoon precipitation, while declining riverflow on the Sutlej from 1991-2004 was attributed to increasing temperatures and subsequent recession of glaciers which subdued runoff from the 1990s maximum. Collins et al (2013), with conflicting arguments, found annual total discharge on the Sutlej upstream at Khab was unrelated to air temperature and declined by 32% from the 1980s to early 2000s, reflecting reduced monsoon precipitation.

In the Jhelum catchment mixed trends of annual runoff were identified from records in excess of 30 years from the 1960's to 1990's (Sharif et al 2013). The northern most tributary of the Jhelum, the Kunhar, had decreasing (non-significant) trends at Naran and Ghari Habibullah, while upward trends were identified on the Jhelum, Neelum, and the Poonch, related to increasing trends of winter precipitation. Conflicting signals of annual riverflow were also observed from long term discharge records from gauging stations in the UIB (Sharif et al 2013, Khattak et al 2011). From ~1970-1997 significant increases of annual riverflow were observed at Kachura on the mainstem Indus (receiving waters from Shyok and Shigar) and at Doyien on the Astore River. Upward trends were also noted at Gilgit and Shyok (Although not significant at Shyok), while decreasing trends were noted at Kharmong and Hunza (Sharif et al 2013). Hunza had the most significant decrease in annual riverflow ($p = <0.05$) out of any station in the UIB (Sharif et al 2013), accredited to high glacierisation, significant cooling of mean summer air temperature, and reduced glacial melt contribution. Increased riverflow from nival catchments (Astore, Gilgit) were reflected by upward trends in winter precipitation (e.g, Archer and Fowler 2004). At Besham Qila an upward (non-significant) trend was observed between 1967 and 1997 (Sharif et al 2013) and from

1967-2005 by Khattak et al (2011), suggesting relatively stable conditions downstream at the outlet of the UIB catchment.

### 4.7.2. Summer riverflow

Undoubtedly summer streamflow is the most important source of water in glacierised Himalayan catchments, accounting for around 80% of total annual riverflow between June-September (e.g, Yu et al 2013), thus trends in summer runoff can be thought to have similar patterns to annual riverflow, but the separation of summer runoff into individual months allows the distinct genetic sources of runoff to be examined. From the glacierised Hunza catchment in the UIB, June-July (snowmelt period) and August-September (glacial melt period) runoff significantly declined from 1966-1997 (Sharif et al 2013), exceeding a 20% reduction from a 1°C fall in temperature since 1961 (Fowler and Archer 2006). Whilst runoff from the glacierised Shyok basin was predicted to reduce by 17% from a 1°C reduction in summer mean air temperature (Fowler and Archer 2006), actual riverflow from 1973-1997 in early (June-July) and late (August-September) summer slightly increased (Sharif et al 2013), despite recorded summer cooling. Runoff increases were also observed from modelled riverflow in the glacierised Shigar catchment (Mukhopadhyay and Khan 2014a). An upward trend in June and July runoff from 1985-2010 was reflected by increased seasonal melting of snow from increasing precipitation. However, August riverflow, which generally reflects glacial melting also increased. This increase was accredited to loss of glacial mass despite summer cooling trends (e.g, Siddique and Hashmi 2012), but the loss of glacial storage was compensated by proportional increase in snow accumulation from monsoon precipitation above >3500m a.s.l. At river gauging stations Kharmong, Kachura and Partab bridge on the main stem Indus River, falling trends in riverflow during the glacial melt period (July-September) were observed from 1995-2009 as a result of falling trends in sensible heat during the same period (Siddique and Hashmi 2012). Likewise, mixed trends were observed during the snowmelt period (May-June), with weak falling trends at Kharmong and Kachura, and a weak rising trend at Partab Bridge.

## 4.8. Glacier fluctuations and mass balance

### 4.8.1. Karakoram

From the early 20th century to the late 1990's majority of glaciers in the Karakoram were observed to be retreating since the Little Ice Age maximum, with the exception of minor advances during the 1970's and 1980's (Mayewski and Jeschke 1979, Hewitt 2005). Karakoram glaciers have complex behaviour with rapid retreat and advance phases, and differences in observed glacier behaviour exist despite similar locations and glacier length (Mayewski and Jeschke 1979, Hewitt 2011). On the whole, Karakoram glaciers declined by around 5% between 1920 and 1960, but glaciers in the 21st century were not at their furthest recorded retreat (Hewitt 2011). Surveys between 1997 and 2002 suggested re-advancement of glacier termini with thickening ablation zones of up to 15m, confined to the highest watersheds in the central Karakoram, postulating the term 'Karakoram anomaly' (Hewitt 2005). Since the publication of the Karakoram anomaly a large number of studies have focused on the health of Karakoram glaciers during the past decade. A variety of techniques including the geodetic method, gravimetric method, glaciological method, 2D areal changes, and surface ice velocities have been used as an attempt to estimate the status of Karakoram glaciers on a region and individual glacier scale.

The geodetic method measures changes in ice elevation by differencing Digital Elevation Models (DEM) or laser altimetry data, and converts mean elevation changes into mass balance by assuming a density of snow and ice loss between 600-900kg $m^{-3}$ (Gardelle et al 2012). Using the geodetic method a positive mass balance over a 5615$km^2$ study area in the central Karakoram was indicated (mean annual mass balance between: +0.05±0.16m $yr^{-1}$ and +0.11±0.22m $yr^{-1}$)(Gardelle et al 2012). Similar observations were made by combining measurements from the Ice, Cloud and land Elevation Satellite (ICESat) and the DEM from the Shuttle Radar Topography Mission (SRTM) from 2003-2009 (Kaab et al 2012). Thickening of glaciers was found in the northern and eastern Karakoram (autumn balance: +0.14±0.06m $yr^{-1}$), whilst glaciers in Jammu and Kashmir, Himachal Pradesh, and central Himalaya declined (Kaab et al 2012). A positive mass balance (excluding surge type glaciers) was also indicated in the western (+0.09±0.15m w.e. $yr^{-1}$) and eastern (+0.12±0.12m w.e. $yr^{-1}$) Karakoram, and also in the Pamir mountain range (+0.12±0.11m w.e. $yr^{-1}$) from 1999-2011, giving credence to the 'Pamir-Karakoram anomaly' (Gardelle et al 2013). Part of the eastern Karakoram (Shyok basin) had area decreases up to 11km$^2$ from 1973-1989 but between 1989-2002 and from 2002-2011 area increases of+8.2±0.33km$^2$ and +5.6±0.21km$^2$ were observed (Bhambri et al 2013), while the neighbouring Chang Chemo river basin had overall no change in area from 1973-2011.

Theoretically, during a period of positive mass balance glacier surface velocities may increase as accumulated snow and ice will be transported to lower elevations as glaciers advance. Based on surface ice velocity estimates using ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) and SPOT (Satellite Pour l'Observation de la Terre) satellite images Scherler et al (2011) found 58% of Karakoram glaciers were stable or slowly advancing at a rate of +8±12m yr$^{-1}$. From 181 glaciers in the Karakoram, glacier speeds increased on average by 5% per decade, indicating a neutral or slightly positive mass balance, while glaciers in the Pamir's decreased their speeds by 43% per decade (Heid and Kaab 2012), but are still linked to a positive mass balance by Gardelle et al (2013). Both regions are well known for surging glaciers which may affect surface ice velocities, but stable, non-surging glaciers, Siachen and Baltoro in the eastern Karakoram have increasing flow speeds between 2001/02-2009/10 and are linked to a positive mass balance (Heid and Kaab 2012). On the contrary, a negative mass balance was reported from Siachen glacier between 1986/87 and 1990/91 (Bhutiyani et al 1999), and continued to retreat 1.9km between 1989 and 2006 (Rasul et al 2008). Increasing winter velocities from non-surging Baltoro glacier since 2003 also provides evidence supporting a positive mass balance in the Karakoram (Quincey et al 2009). This is supported by Mayer et al (2006) who suggested mass balance of Baltoro was close to zero. Although gains in glacier mass have been inferred for Karakoram glaciers, glacier mass balance of the Indus basin, for which a large proportion of the Karakoram lies within, is largely negative. This is because glaciers which have mass gains are in basins draining north of the Indus basin and do not contribute to the glacier imbalance of the Indus (Kaab et al 2012; supplementary information).

### 4.8.2. North-west Himalaya

Glaciers of the North-western Himalaya generally reached their maximum extent between 1850 and 1880 (Mayewski and Jeschke 1979), and following their retreat phase in the 20$^{th}$ and 21$^{st}$ century deglaciation varied considerably between regions due to topographic and climatic influences (e.g, Mayewski and Jeschke 1979, Kulkarni et al 2007). Glaciers belonging to the Nanga Parbat massif generally had minor retreat rates(6m $yr^{-1}$) between 1860-1960, with the exception of Raikot and Chungphar glaciers which had retreat rates up to 30m $yr^{-1}$ between 1930 and 1950 (Mayewski and Jeschke 1979). In the 1980's a re-advance of Raikot glacier terminus was observed (Gardner 1986, Schmidt and Nusser 2009), but glacier retreating continued into the 21st century (Schmidt and Nusser 2009). To the east, glaciers of the Lahaul-Spiti, Himachal Pradesh region, had larger retreat rates, varying from 20.5m to 62.5m $yr^{-1}$ since the Little Ice Age maximum (Mayewski and Jeschke 1979). Between 1999 and 2004 significant thinning rates of Lahaul-Spiti glaciers were recorded at low elevations (up to 10m below 4400m a.s.l), with mass balance of a 915km$^2$ area between -0.7 and -0.85m w.e. $yr^{-1}$ (Berthier et al 2007), which was twice the recorded ice loss in the central Himalaya (-034 m w.e. $yr^{-1}$) (Dyurgerov and Meier 2005 in Berthier et al 2007)). More recently however, it is believed ice wastage over the past 22 year (1988-2010) in Lahaul-Spiti has been moderate (Vincent et al 2013), with mass balance estimates of -0.44±0.09m w.e. $yr^{-1}$ between 1999 and 2011 for a region covering 2110km$^2$ of glaciers. From combined direct field measurements and estimates of mass balance using the geodetic method it was inferred mass balance was slightly positive or near neutral (+0.09±0.15m w.e. $yr^{-1}$) between 1988 and 1999 (Vincent et al 2013), similar to observations at Raikot glacier (Schmidt and Nusser 2009).

In the Himachal Pradesh region, glacier losses were substantial in the late 20th century (e.g, Kulkarni 1992, Kulkarni et al 2004, Wagon et al 2007, Kaab et al 2012). From 1962-2001 a glacierised area of 2077km$^2$ with 466 glaciers declined by 21% to 1628km$^2$ (Kulkarni et al 2007). For individual basins, losses were similar, ranging from 19% glacier area losses in the Baspa basin (tributary of the Sutlej), 21% in the Chenab basin, and 22% in the Parbati basin (Kulkarni et al 2007). Further north, in the Ladakh range, glaciers of the Kang Yatze massif below the Indus River at Leh, declined by 14.3% from 1962-2010, with smaller glaciers being most affected (Schmidt and Nusser 2012).

# 5. Methods

## 5.1. Meteorological and discharge data

Criteria for selecting meteorological data were based on length and continuity of time series record. Ideally, records should be available at monthly resolution, be homogenous, and be located at High altitudes close to glaciers to enable indication of climatic variation since the Little Ice Age maximum. Data from meteorological stations in the UIB were obtained directly from the Pakistan Meteorological Department (PMD), while data in the Himalaya and surrounding areas were obtained by a combination of the Royal Netherlands Meteorological Institute (KNMI), Indian Meteorological Department (IMD), and the Bhakra Beas Management Board (BBMB).

Climate stations in the UIB were located at altitudes between 455-3015m a.s.l, stations at higher elevations generally having shorter records. Station locations and characteristics can be seen in figures 3-1 – 3-5 and tables 5-1 and 5-2. Records were generally 50+ years in length with longer records of around 100 years being available at Gilgit and Skardu. Records at Gilgit and Skardu are not strictly homogenous as site changes occurred after 1947, moving a distance of 1.5km and 8km, respectively, from the British administrative headquarters to their respective airports (Archer and Fowler 2004). Climate stations in the western-central Himalaya were particularly sparse at high elevations. Records were mainly chosen from climate stations at low elevations south of the main Himalaya with most records ending in the mid/late twentieth century and thus may not be entirely representable of conditions within a basin. Short but recent records of precipitation were available at Bhakra and Rampur. Climate stations at Sialkot, Amritsar and Ludhiana are located in cities, thus temperature records may be influenced by the growth of industrialisation from the late nineteenth century. Long uninterrupted records of precipitation were however available at district level in the western Himalaya courtesy of the IMD between 1900 and 2013. Individual climate stations are preferred over district series, as district data is the arithmetic mean of total monthly precipitation from individual stations within the district and can be interchangeable from year-to-year if a particular station is neglected from the series. Climate stations along the Tibetan Plateau were situated at distances far from gauging stations but were examined to view climatic variation at the source of the Indus and Sutlej.

Climatic parameters were chosen on their glaciological and hydrological influence. Winter precipitation (P11-5) was calculated from the sum of monthly totals of precipitation from November (year n-1) to May (year n). Precipitation accumulates as snow at high altitudes during winter and nourishes glaciers after summer melting but also adds a significant contribution to runoff in spring and early summer as temperature and radiation increases. Summer precipitation, which includes both monsoon and non-monsoon sources, was calculated from the sum of monthly total precipitation between June and September (P6-9). Summer precipitation can have negative or positive influences on runoff depending on the proportion of ice within a basin. Monsoon adds a significant component of riverflow in the Himalayas, but reduces runoff in glacierised basins as precipitation falling as snow has a higher albedo than underlying ice. Average summer air temperature from May to September (T5-9) indicates energy availability to melt snow and ice and has high correlations with runoff from glacierised basins (e.g. Collins 1987). T5-9 was calculated from average monthly temperatures between May and September at all stations except for stations in the UIB where T5-9 was calculated from average monthly minimum and maximum temperatures, and is therefore a poor substitute. If a data value for a climatic parameter was missing for a particular month then the entire year was omitted from the time series.

**Table 5-1: Rain gauge stations and characteristics of seasonal precipitation from stations in the UIB and Himalayas.**

| Station | Period of record | Elevation (m) | Mean P6-9 (mm) | Mean P11-5 (mm) | % of P6-9 to total annual precipitation | % of P11-5 to total annual precipitation | Data source |
|---|---|---|---|---|---|---|---|
| Gilgit | 1892-2013 | 455 | 45.7 | 81.9 | 34.4 | 61.7 | PMD |
| Skardu | 1900-1999 | 2200 | 40.5 | 143.3 | 22.3 | 76.6 | PMD |
| Bunji | 1952-2013 | 1250 | 59 | 87 | 37.9 | 56 | PMD |
| Astore | 1954-2013 | 2000 | 96.4 | 372.4 | 19.5 | 75.4 | PMD |
| Chilas | 1953-2013 | 1075 | 42.3 | 139.4 | 23.2 | 73 | PMD |
| Srinagar | 1893-2013 | 1587 | 197.6 | 449.4 | 28.9 | 65.6 | KNMI |
| Srinagar district | 1902-2010 | | 237.3 | 671.1 | 25 | 70.8 | IMD |
| Ludhiana | 1868-1990 | 255 | 556.6 | 146.2 | 78.1 | 20.5 | KNMI |
| Bhakra | 1983-2012 | | 828.4 | 303* | 73 | 26.7 | BBMB |
| Shimla | 1863-1987 | 2202 | 1167 | 352.1 | 76.2 | 23 | KNMI |
| Shimla district | 1901-2012 | | 797.2 | 378.9 | 66.1 | 31.4 | IMD |
| Dehradun district | 1901-2013 | | 1748.1 | 271.7 | 84.6 | 13.4 | IMD |
| Chakrata | 1901-1967 | 581 | 1239.1 | 391.8 | 77 | 24.3 | KNMI |
| Rampur | 1999-2004 | | 431.4 | 263.5 | 58.2 | 35.6 | BBMB |

Discharge data was selected on the same criteria as meteorological data. It was important to obtain discharge data from glacierised basins to infer state of glaciers, but it was equally as important to obtain data downstream at large dams to examine long term water resources. List of gauging stations and characteristics of basin areas can be seen in table 3-1. Data of riverflow in the Indus, Jhelum and Chenab was measured by the Water and Power Development Authority (WAPDA). Discharge in the UIB is based on manual measurements, calculated by measuring stage and converted to discharge using rating curves. Measurements take place at hourly intervals during the summer ablation season and once daily during the winter, and the accuracy of the data are described by Archer (2003) as 'moderate quality at best'. Measurements of discharge at Mangla dam since the first impounding are calculated indirectly from the water level in the reservoir. Riverflow in the Sutlej was available at several gauging stations along the mainstem river and was made available by the BBMB and DHI: water and environment.

**Table 5-2: Air temperature stations in the UIB and Himalaya.**

| Station | Period of record | Elevation | Mean T5-9 | Data source |
|---|---|---|---|---|
| Gilgit | 1961-2013 | 455 | 24.1 | PMD |
| Askole | 2007-2009 | 3015 | 14.8 | |
| Bunji | 1961-2013 | 1250 | 26.2 | PMD |
| Astore | 1962-2013 | 2000 | 18 | PMD |
| Chilas | 1961-2013 | 1075 | 30 | PMD |
| Hotan | 1942-2013 | 1375 | 23.9 | KNMI |
| Kashi | 1959-2011 | 1291 | 22.5 | KNMI |
| Shiquanhe | 1961-2013 | 4279 | 10.1 | KNMI |
| Srinagar | 1893-2011 | 1587 | 21.4 | KNMI |
| Jhelum | 1947-2013 | 234 | 30.8 | KNMI |
| Sialkot | 1964-2013 | 256 | 30.6 | KNMI |
| Amritsar | 1949-2009 | 234 | 30.2 | KNMI |
| Ludhiana | 1876-1983 | 247 | 31.4 | KNMI |
| Shimla | 1876-1987 | 2202 | 18.1 | KNMI |
| Mukteshwar | 1897-2012 | 2311 | 17.6 | KNMI |

Annual totals of discharge are generally shorter in length in comparison with climatic data, data commencing in the 1960s and 1970s for gauging stations in the Indus and Sutlej. Longer records are available at Mangla and Bhakra, dating back to 1923 and 1920, respectively, but end in the early 2000s. Annual totals of discharge from January to December (Q1-12) were used instead of individual months as not all gauging stations had discharge data available at monthly resolution. Annual discharge at Bhakra was the sum of discharge from June to May (Q6-5). Construction of dams at Tarbela (1977), Mangla (1967), Marala (1968) and Bhakra (1963) may have affected homogeneity of discharge records. Riverflow at Bhakra is artificially enhanced by waters diverted from the Beas-Sutlej link canal after 1977, where waters enter the Sutlej below Suni at Dehar power station. Flows upstream are natural, dams do exist but are run-of-river type dams which do not impound water. Discharge records at Khab ceased since June 2005 as the gauging station washed away during a flood event.

## 5.2. Basin areas and basin characteristic

Basin areas, defined as the drainage area upstream of a gauging station, were automatically delineated using data from The Shuttle Radar Topography Mission (SRTM) (2003) at 90m resolution with the help of ArcGIS 9.3.1. ArcGIS spatial analyst hydrology tool was the chosen method to extract basin areas. After mosaicing several individual SRTM DEM data sets into one large STRM the first step was to use the fill tool to remove any imperfections (sinks) within the Digital Elevation Model. Flow direction was calculated to indicate the direction of water movement from cell-to-cell before calculating the accumulation of flow down slope. Theoretically it is possible to compute the basin area at any given point of a river after applying the Flow direction and Flow accumulation. Basin areas used in this study can be seen in table 3-1, and a comparison of calculated basin areas used in this study with previously published basin areas can be seen in table 5-3. Catchment areas for basins in the UIB were closely matched with several authors (table 5-3), the highest discrepancy being 616km$^2$ between Hunza (this study) and Hunza (WAPDA). Best estimates for basin areas at Besham and Tarbela by Khan et al (2014) were used as they identified several critical points which were not part of the Indus catchment. The basin area at Bhakra had a difference of 4651km$^2$ (around 8%) compared with Singh and Jain (2002), but lack of previously published basin areas in the Sutlej along with high accuracy of extracted catchment areas in the UIB led to believe the Sutlej basin areas were of reasonable accuracy.

Total ice area of a catchment displayed large variance between different authors (table 5-4). There was no specific criterion to determine which the most accurate source was. Ice areas by Yu et al (2013) were chosen based on being the most up to date record and most consistent. Ice areas for Tarbela and Sutlej were selected from ICIMOD (2012). Total ice area of a basin was the preferred parameter over percentage glacierisation, particular in the UIB where the lower ice-free valley bottoms are effectively hydrologically inactive as they receive little or no precipitation to contribute to runoff.

Total average seasonal and annual precipitation of a basin, described in chapter 3, was calculated from gridded datasets of total monthly average precipitation between 1950 and 2000 obtained from Worldclim.org. Datasets were mosaiced in ArcGIS and extracted from the basin shapefiles calculated using the spatial analyst hydrology tool.

**Table 5-3: Comparison of basins area (km²) with this study and previously published authors.**

| River | Station | this study | Sharif et al (2013) | Yu et al (2013) | WAPDA (in Kan et al 2014) | Khan et al (2014) | Archer and Fowler (2008) | Singh and Jain (2002) |
|---|---|---|---|---|---|---|---|---|
| Shyok | Yogo | 33,117 | 33,350 | | 33,670 | 33,143 | | |
| Hunza | Dainyor | 13,773 | 13,925 | 13,732 | 13,157 | 13,732 | | |
| Gilgit | Gilgit | 12,681 | 12,800 | 12,680 | 12,095 | 12,671 | | |
| Astore | Doyian | 3,900 | 3750 | 3988 | 4040 | 3899 | | |
| Indus | Besham | | 166,096 | 166,096 | 162,393 | 164,853 | | |
| Indus | Tarbela | | | | 166,019 | 172,159 | | |
| Jhelum | Mangla | 33,468 | | | | | 33,342 | |
| Chenab | Marala | 26,298 | | | | | | |
| Sutlej | Khab | 41,278 | | | | | | |
| Sutlej | Rampur | 45,907 | | | | | | |
| Sutlej | Luhri | 47,807 | | | | | | |
| Sutlej | Suni | 48,857 | | | | | | |
| Sutlej | Bhakra | 52,224 | | | | | | 56,875 |

**Table 5-4: Comparison of Ice areas (km²) between previously published authors.**

| catchment | Campbell (2005) | Akhtar et al (2008) | ICIMOD (2012) | Yu et al (2013) |
|---|---|---|---|---|
| Shyok | 3547.84 | | 5937.7 | 6221 |
| Hunza | 4677.34 | 4688 | 2753.9 | 4339 |
| Gilgit | 968.1 | 915 | 938.3 | 994 |
| Astore | 607.03 | 612 | 239.6 | 450 |
| Besham | | | | 14,889 |
| Tarbela | | | 15,061.7 | |
| Jhelum | 148.18 | | 222.8 | 0 |
| Chenab | | | 2341.2 | 2708 |
| Sutlej | | | 1315 | |

## 5.3. Statistical analysis

Year-to-year variability of annual runoff and precipitation was calculated for several periods of differing but uniform lengths. The variability of annual runoff and precipitation from year-to-year can be estimated using the coefficient of variation (CV):

$$Cv = \frac{\sigma}{\overline{Q}}$$

Where $\sigma$ is the standard deviation and $\overline{Q}$ is the arithmetic mean of total annual discharge over a specific period. Uncertainty of CV can arise from short sampling periods where CV reflects a shift in climatic regime. It is recommended a sampling period of 30-35 years of uninterrupted, homogenous records should be adopted to reflect best accuracy (Moore 1992, Tollan 1972). A short 21 year period from 1980-2000 was adopted to utilise all gauging stations. Longer periods between 25 and 43 years were used to analyse select gauging stations where length of records were longer. CV of seasonal precipitation, P6-9, P11-5, and total annual precipitation from January to December (P1-12) were also calculated for the respective periods of discharge to help provide explanation for CV of Q1-12.

Correlation analysis for common periods of time series records was used as a statistical measure to examine the relationship between climatic variables and discharge. It was important to understand how seasonal climate related between neighbouring stations and between stations over large distances in regions with opposing climatic regimes. Correlations were examined between totals of seasonal precipitation (P6-9 and P11-5) and average summer air temperatures (T5-9) to see whether precipitation influenced energy availability for melting of snow and ice the following summer. Most importantly climate-runoff relationships were established and displayed against the ice cover of a basin. Long correlations in excess of around 40-50 years of overlap between time series are preferred over short overlapping records as short correlations may occur due to chance. Pearson's product moment correlation analysis was the preferred method of correlation over Spearman's rank correlation which is largely used for the analysis of non-parametric data.

# 6. Results

## 6.1. Long term year-to-year fluctuations of total summer precipitation

### 6.1.1. UIB

Year-to-year fluctuations of total summer precipitation displayed similar evolutions at all stations in the UIB (figure 6-1). Lowest annual totals of summer precipitation were recorded at Skardu (average P6-9 of 40.5mm), accounting for on average 22.3% of total annual precipitation (table 5-1). With large year-to-year fluctuations, P6-9 at Skardu declined from the wettest decade in the 1900s, from the series maximum in 1909 (127.6mm), to relatively dry years in the 1920s and 1930s. With some recovery in the 1950s precipitation returned to dry years in the 1960s before inclining to wetter years in the 1990s, matching similar levels of the 1950s (figure 6-1). At Gilgit (average P6-9 of 45.7mm, around 34.4% of total annual precipitation) in the northwest UIB total summer precipitation displayed no specific trend from the 1890s-1980s. Decreasing in the 1960s P6-9 increased substantially from around the 1980s-2000s by around 36% with the wettest individual year in 2010 (146.6mm). Total quantity of summer precipitation increased on the border of the western Himalayas at Astore (average P6-9 of 96.4mm) and Bunji (average P6-9 of 59mm) and accounted for on average 19.5% and 39.7% of total annual precipitation. Both displayed similar year-to-year variations, declining in the 1960s before inclining to the wettest period in the 1990s, increasing by around 72% and 103% from the 1960s-1990s. Precipitation dipped slightly in the 2000s at Astore and Bunji, but still remained the second wettest decade of the series after 1990 at both stations. Further south, on the Indus at Chilas, P6-9 was also down in the 1960s, often failing to reach above 20mm annually, but increased from the 1970s by around 113% to the 1990s, with the series maximum in 1992 (152.7mm). P6-9 reduced substantially in 1998 and remained below the series average from 2000-2005 before making some recovery toward the end of the series.

### 6.1.2. Western Himalayas

Total summer precipitation at Srinagar was marginally higher than those climate stations situated in the UIB (average P6-9 of 197.6mm), accounting for on average

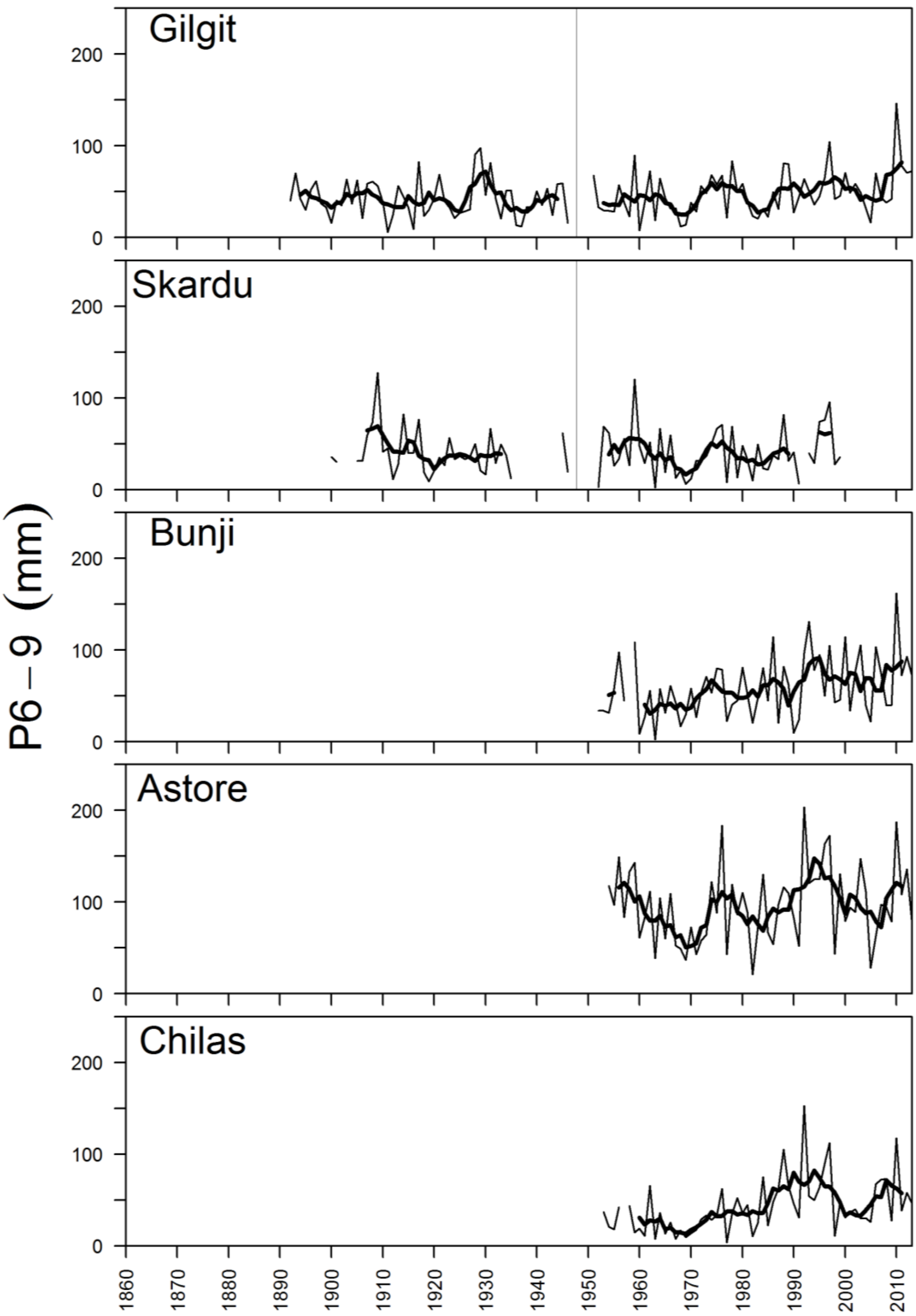


**Figure 6-1: Year-to-year annual totals (thin line) and five-year moving average (thick line) of total summer precipitation from June-September (P6-9) at UIB climate stations; Gilgit, Skardu, Bunji, Astore, and Chilas. Vertical line indicates site changes at Gilgit and Skardu after 1947.**

around 30% of total annual precipitation (table 5-1). Summer precipitation varied considerably in the late 19th and early 20th century (figure 6-2), declining from the early 1900s maxima to a minimum in 1934 (66mm), rising to the 1950s, before falling below the series average in the 1960s and 1970s. Precipitation increased by 47% from the 1960s to the 1990s, and by 34% from the 1960s-2000s, similar to the rise in precipitation from climate stations in the UIB, but still failed to exceed the early 1900 maxima. Precipitation recovered slightly in the early 2000s, climbing to the series maximum in 2006 (401mm). The district of Srinagar series (average P6-9 of 237.3mm) generally followed a similar pattern to the Srinagar series, declining from the 1900s, increasing to the 1950s and then declining in the 1960s. Unlike Srinagar, summer precipitation followed no specific trend from 1960s, declining by only 2% from the 1960s-2000s, decreasing from the late 1990s. On the Punjabi plains below Bhakra dam, summer precipitation at Ludhiana (average P6-9 of 557mm) accounted for on average over ¾ of total annual precipitation but was two-three times less than average total P6-9 to the east at Chakrata and Dehradun (table 5-1). Summer precipitation at Ludhiana varied considerably from the 1860s to the 1940s with no specific trend. Slightly down in the late 1930s/ early 1940s precipitation increased to the 1960s with maximum quinquennia centred on 1964. Precipitation declined to the 1970s but recovered in the 1980s. At Bhakra (average P6-9 of 828mm) summer precipitation increased slightly from the mid-1980s to late 1990s but decreased slightly in the early 2000s.

Long term monsoon precipitation from meteorological stations at slightly higher elevations near the Sutlej basin varied cyclically from the 19th and 20th century (figure 6-3), falling considerably since the 1950s at Shimla and since the 1960s at Dehradun and Shimla district (figure 6-3). Precipitation remained down in the last 1/3 of the 20th century despite a slight rising trend in the 1980s/1990s, falling by 20-30% at Shimla district and Dehradun district since 1960/69-1990/99. From the 21st century, precipitation at Dehradun district increased sharply by around 60% from the 1980s, attaining similar levels of the wet 1920s, whilst precipitation at Shimla district recovered to similar levels attained during the 1970s.

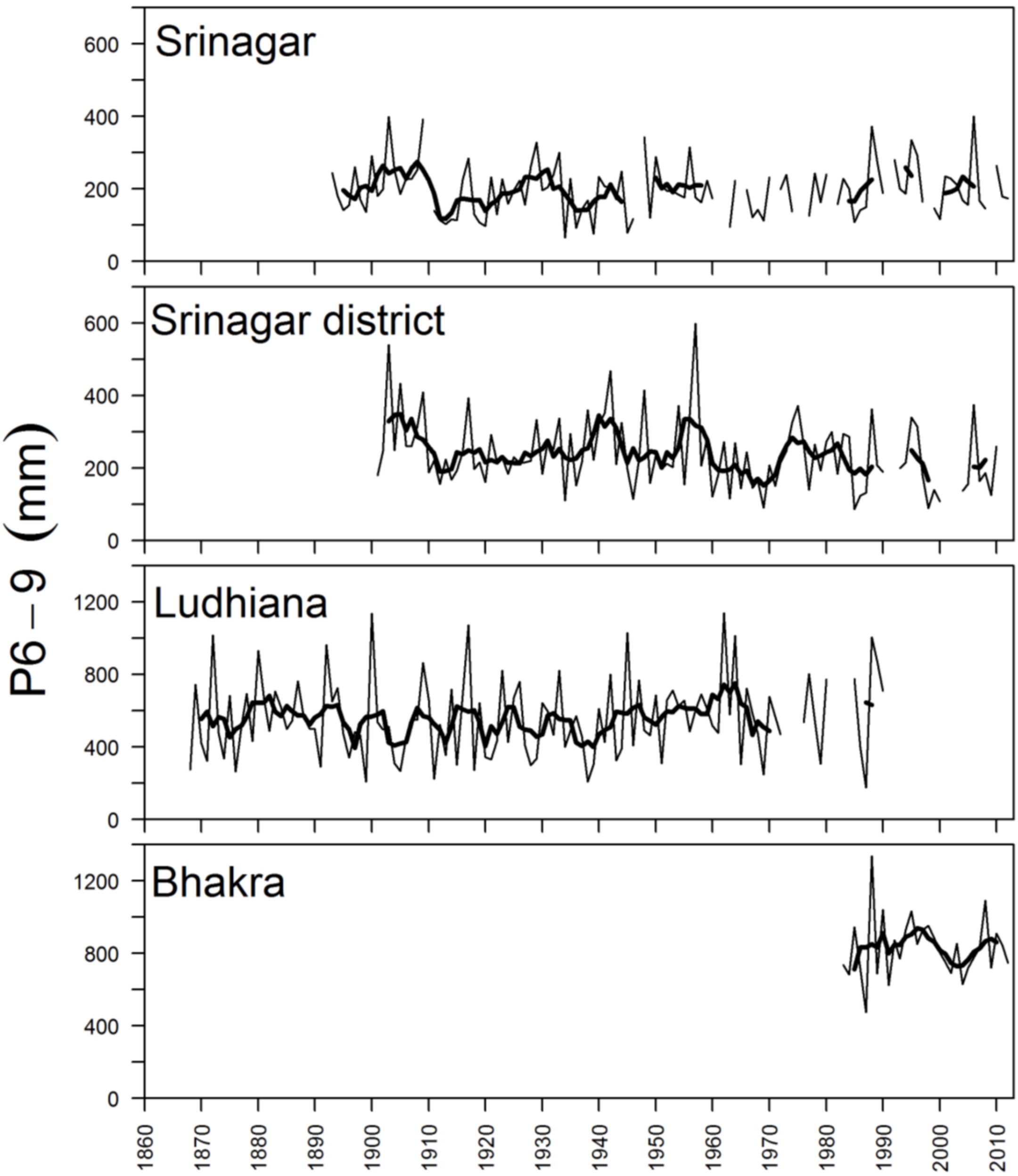


**Figure 6-2: Year-to-year annual totals (thin line) and five-year moving average (thick line) of total summer precipitation from June-September (P6-9) at western Himalayan climate stations; Srinagar, Srinagar district, Ludhiana and Bhakra. Note that precipitation ranges differ between plots.**

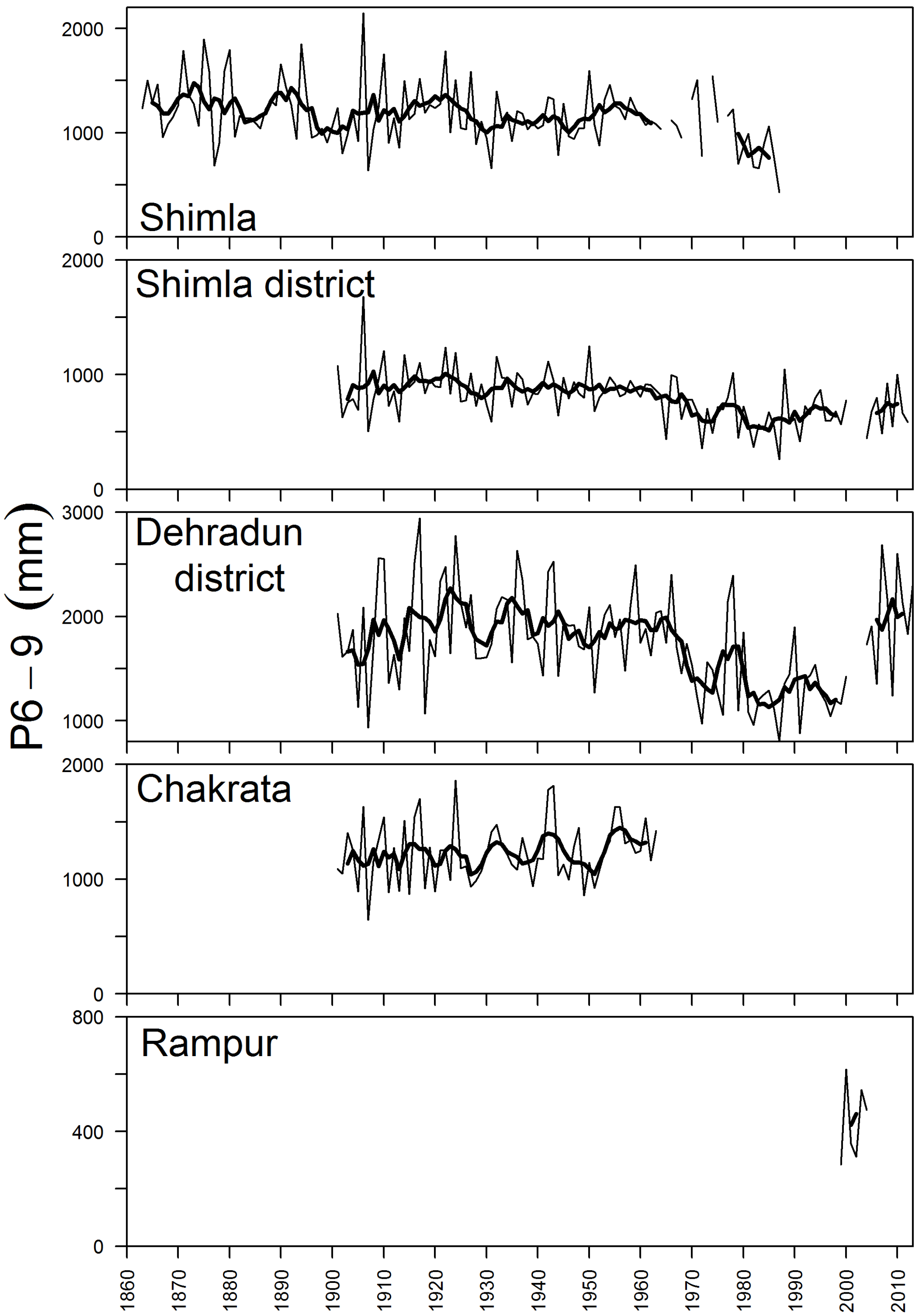


**Figure 6-3: Year-to-year annual totals (thin line) and five-year moving average (thick line) of total summer precipitation from June-September (P6-9) at western Himalayan climate stations; Shimla, Shimla district, Dehradun district, Chakrata and Rampur. Note that precipitation ranges differ between plots.**

## 6.2. Long term year-to-year fluctuations of total winter precipitation

### 6.2.1. UIB

Total winter precipitation from November to May varied considerably across the UIB from the late nineteenth to the early twenty-first century (figure 6-4). Total quantity of winter precipitation was least at Gilgit (average P11-5 of 81.9mm) and accounted for on average 61.7% of total annual precipitation (table 5-1). From year-to-year P11-5 at Gilgit varied greatly; declining in the early 20$^{th}$ century, inclining to the wet 1920s and 1930s (average decadal precipitation ~92 and 99.1mm), and then fluctuating around the series average in the 1970s and 1980s before inclining slightly to the 1990s and 2000s, but remaining drier than the 1920s and 1930s. Further east at Skardu airport (average P11-5 of 143.3mm, 76.6% of total annual precipitation) winter precipitation declined in the 1920s and 1930s but increased to the early 1970s following a broken series in the 1940s and 1950s. After a slight decrease in the late 1970s/early 1980s precipitation increased from around~160mm to ~200mm in the wettest decade of the 1990s, with the series maximum in 1996 (359.7mm). P11-5 at Astore (average P11-5 of 372.4mm, 75.4% of total annual precipitation) was almost three times greater than at Skardu being situated on the border of the western Himalaya but had similar evolutions, precipitation rising from the 1980s to maxima in the 1990s. From the series maximum in 1996 winter precipitation declined in the early twenty-first century to a record minimum in 2008 (93.8mm), with the decade average of 2000-2009 falling below the series average by around 50mm. Precipitation at Chilas (average P11-5 of 139.4mm, 73% of total annual precipitation) increased from the late 1970s, rising steadily to a series maximum in 1996 (480.3mm), before falling to similar levels of the 1960s in the 2000s. P11-5 at Bunji (average P11-5 of 87mm, 56% of total annual precipitation) increased from around 70mm in the 1960s to around 85mm in the 1990s but displayed large year-to-year variations, particularly in the 2000s where the series minimum and series maximum were recorded in 2001 and 2003, respectively.

### 6.2.2. Western Himalaya

In the Western Himalaya, total winter precipitation was greatest at Srinagar (mean P11-5 of 449.4mm), accounting for on average 65.6% of total annual precipitation (table5-1). Precipitation generally followed an increasing trend from the 1890s, rising to the wettest decade in the 1960s, falling in the 1970s, before inclining by around 43% to the

1980s with particular wet years in 1983 and 1986. Slightly above the series average in the 1990s, precipitation declined from 1996 and remained down in the 2000s, declining by around 19% from the 1960s (figure 6-5). Year-to-year fluctuations of P11-5 at Srinagar district (average P11-5 of 449.6mm, 70.8% of total annual precipitation) followed a contrasting pattern to Srinagar in the first half of the 20th century, winter precipitation declining considerably to the 1950s. From the 1950s, precipitation followed a similar pattern to Srinagar, declining to the dry 1970s, increasing in the 1980s, and then falling in the late 1990s/ early 2000s. Temporal variation of P11-5 at Ludhiana (average P11-5 of 146.2mm) varied considerably from year-to-year on an underlying declining trend from the 1900s to a minimum in the 1920s, and 1960s. Precipitation increased from the 1970s, similar to the rise of P11-5 at Srinagar and Srinagar district, with the record terminating in the 1980s (figure 6-5). The relatively short record at Bhakra (average P10-5 of 303mm, 26.7% of total annual precipitation) declined from the 1980s to 2011, reaching a minimum in 2004 (117mm).

At higher elevations, winter precipitation at Shimla (average P11-5 of 352.1mm, 23% of total annual precipitation), varied greatly on an underlying decreasing trend from the late 19th and early 20th century maxima, to a minimum in 1932 (126.2mm), declining from around ~400mm in the 1900s to ~300mm in the 1940-70s with recovery in the 1980s (figure 6-6). The Shimla district series (average P11-5 of 378.9mm, 31.4% of total annual precipitation) showed a similar pattern to that of Shimla but displayed a slight increasing trend from the 1940s to 1960s. The 1970s were dominated by a period of drought but was offset by a large increase during the 1980s, before falling in the 2000s to similar levels of the 1940s. A similar pattern was displayed at Chakrata and Dehradun district, both falling from the 1900s, with Dehradun district falling to a minimum in the 1970s. From the 1970s precipitation increased substantially by around 60% to the 1980s, with quinquennia maximum centred on 1981 and the series maximum in 1982 (738.9mm), but fell in the 1990s and 2000s to similar levels of the 1950s (figure 6-6).

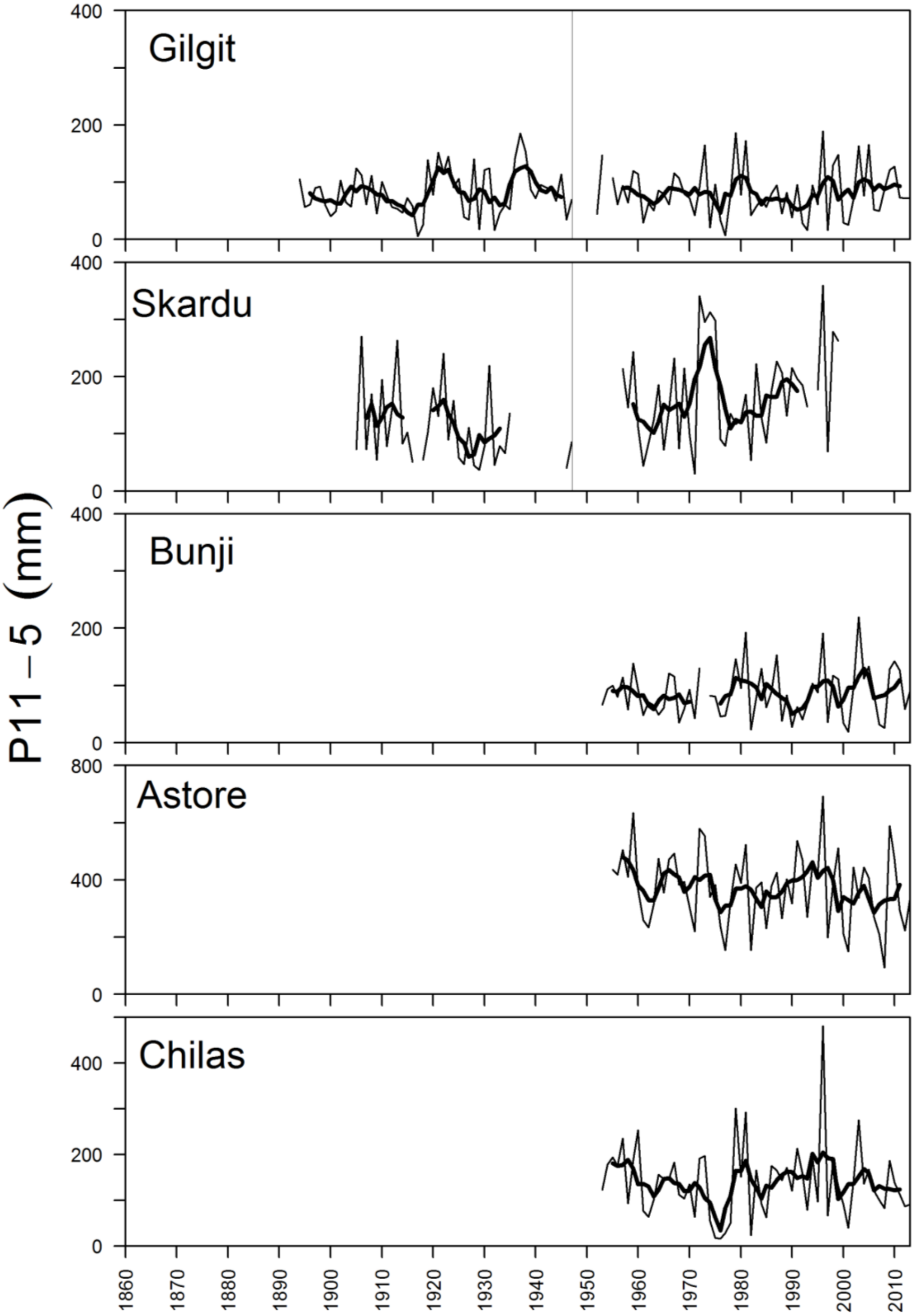


**Figure 6-4: Year-to-year annual totals (thin line) and five-year moving average (thick line) of total winter precipitation from November to May (P11-5) at UIB climate stations; Gilgit, Skardu, Bunji, Astore, and Chilas. Note that precipitation ranges differ between plots. Vertical line indicates site changes at Gilgit and Skardu after 1947.**

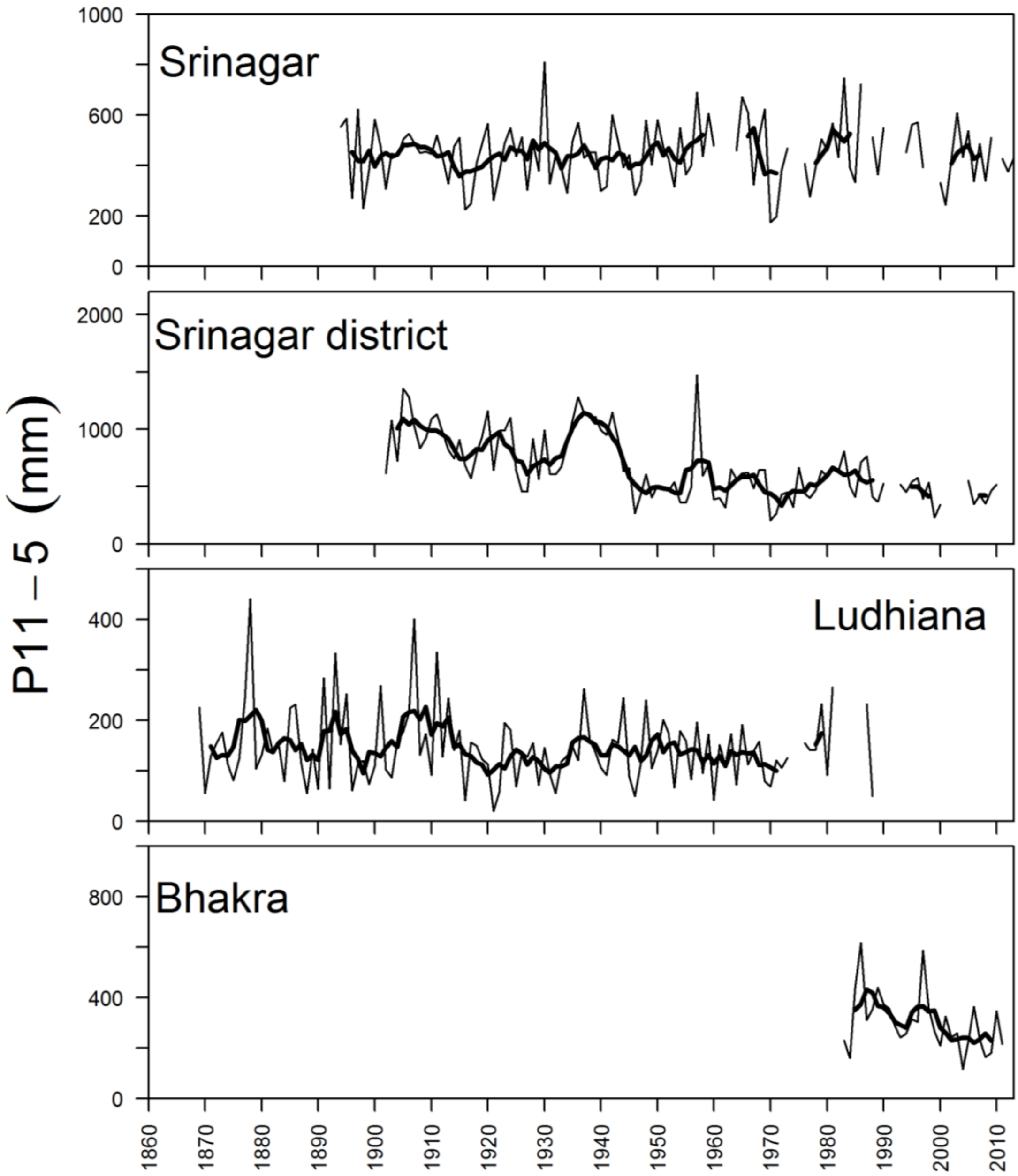


**Figure 6-5: Year-to-year annual totals (thin line) and five-year moving average (thick line) of total winter precipitation from November to May (P11-5) at western Himalayan climate stations; Srinagar, Srinagar district, Ludhiana, and Bhakra. Note that precipitation ranges differ between plots.**

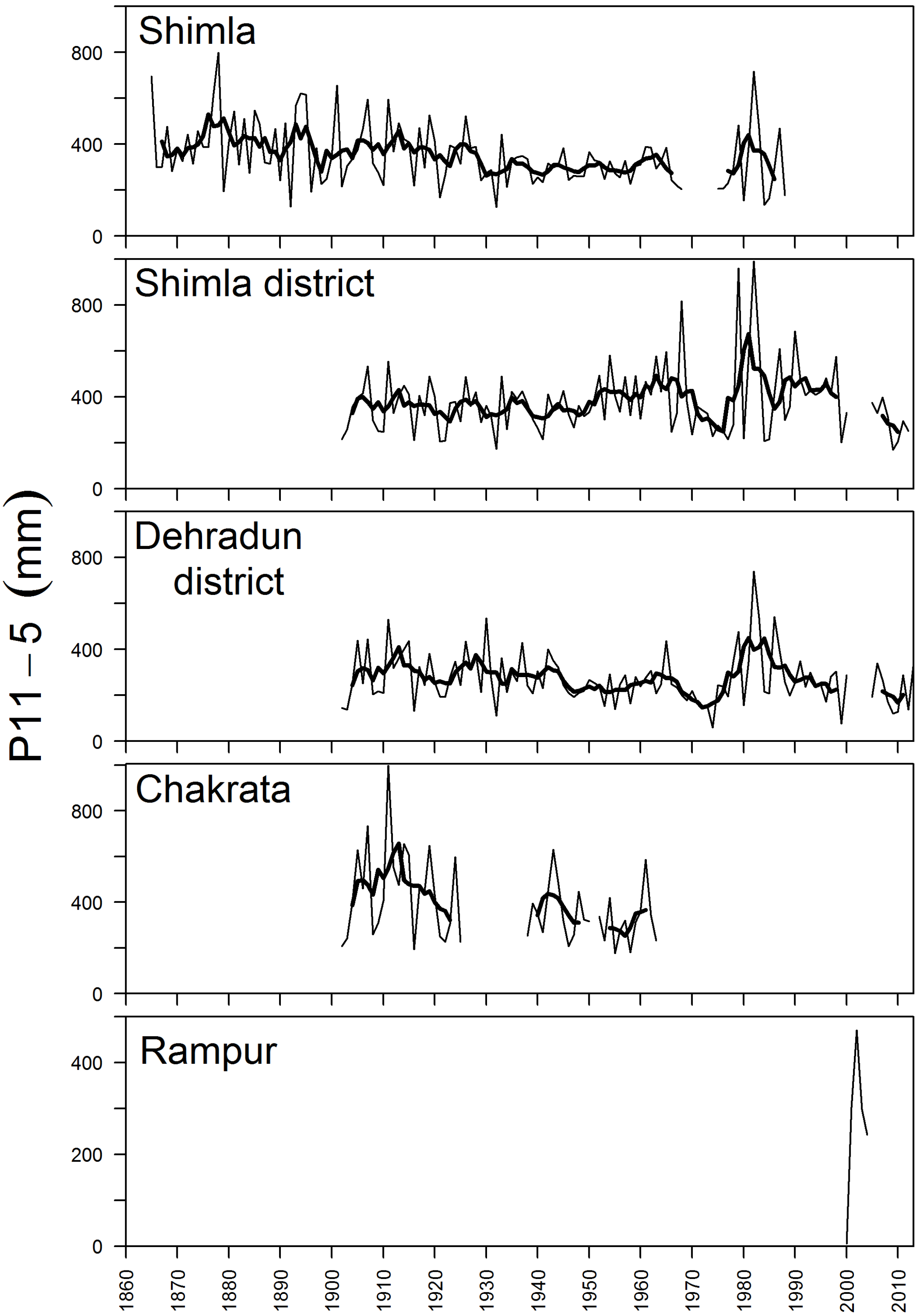


**Figure 6-6: Year-to-year annual totals (thin line) and five-year moving average (thick line) of total winter precipitation from November to May (P11-5) at western Himalayan climate stations; Shimla, Shimla district, Dehradun district, Chakrata, and Rampur. Note that precipitation ranges differ between plots.**

## 6.3. Long term year-to-year fluctuations of average summer air temperature

### 6.3.1. UIB

Mean summer air temperature from May to September provides an indication of energy availability for melting of snow and ice. Generally, year-to-year fluctuations of T5-9 declined or displayed no specific trend in climate stations in the UIB since commencement of records began in the 1960s (figure 6-7). A reduction in temperature was most notable in the valley of the Indus at Bunji where mean summer air temperature declined on average by 1.5°C from the 1960s-2000s, summer temperatures declining steeply from the series maximum in 1961 (29.55°C) to a minimum in 1989 (22.94°C). Year-to-year variations of T5-9 were mirrored at Gilgit and Chilas, both declining from the warmer 1970s, with warm years in 1971, 1973 and 1977, to slightly cooler years in the late 1980s, Chilas declining considerably in the 2000s, declining to a minimum in 2010 (26.38°C). At Astore, summer air temperature declined from the 1970s maxima (series maximum of 13.36°C in 1973) to a minimum in 1989 (9.22°C) but recovered from the 1990s-2000s to similar levels of the warm 1970s, declining by ~0.08°C.

### 6.3.2. Tibet

Long term annual fluctuations of summer air temperature on the Tibetan Plateau varied considerably from year-to-year (figure 6-8). Temperatures at Hotan declined from the 1950s to cooler years in the 1960s before returning to warmer years in the 1970s. Dipping slightly in the 1980s and 1990s, summer temperatures increased markedly to the 2000s, rising by around 1°C from the 1980s-2000s. Temporal variations of T5-9 at Hotan were mirrored at Kashi; cool in the 1960s, late 1980s and early 1990s, warm in the 1970s, but at its maxima in the 2000s. Temperatures increased by around 1.5°C from the 1980s-2000s reaching the series maximum in 2008 (24.6°C) followed by the second warmest summer in 2011 (24.3°C). T5-9 at Shiquanhe was down in the 1960s but increased from the 1980s- with interrupted records- to its maxima in 2007 (12.0°C).

### 6.3.3. Western Himalaya

Mean summer air temperature at Srinagar followed a surprisingly similar pattern to climate stations in the UIB and Tibet; declining from a maxima in the 1970s, with the

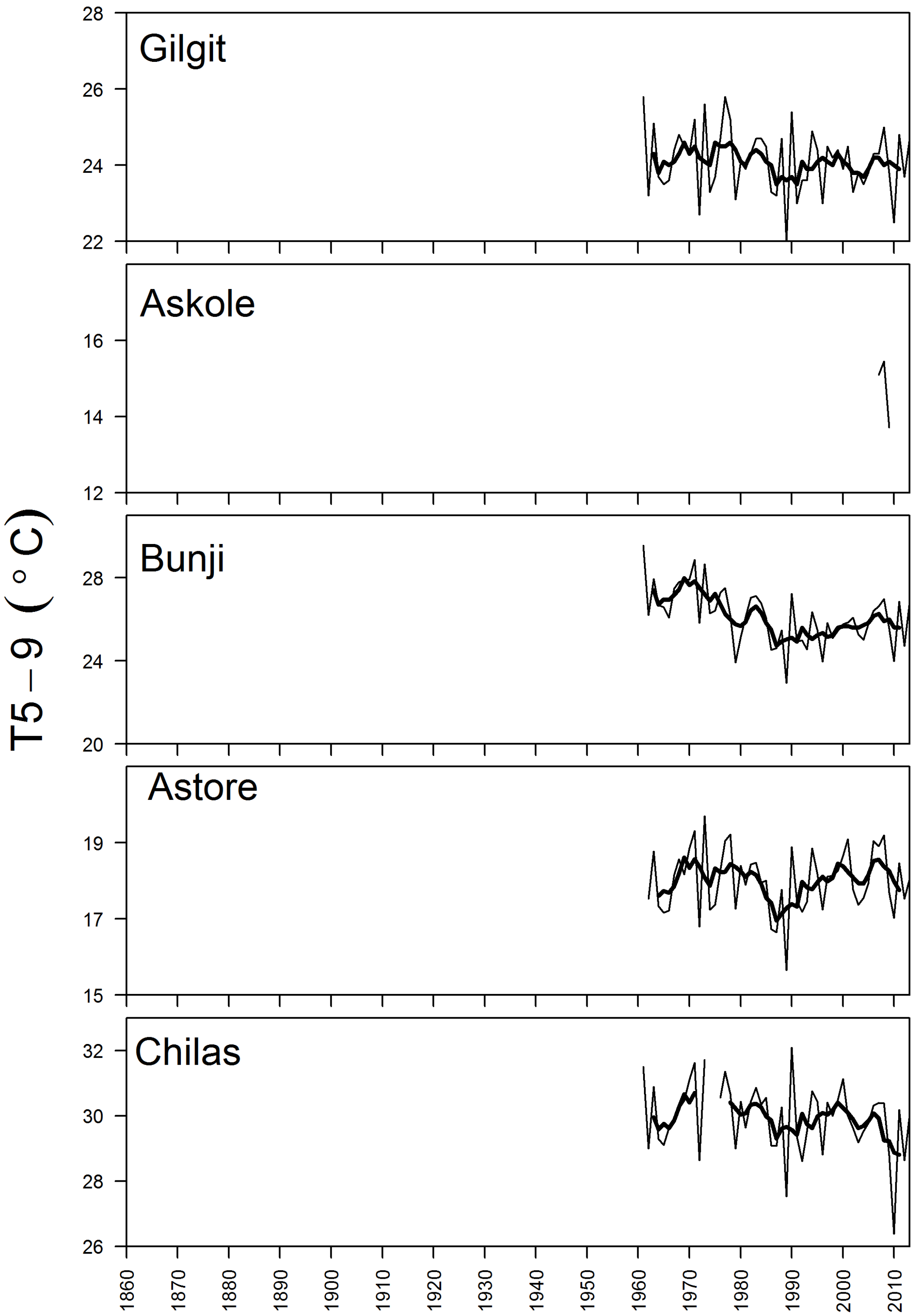


**Figure 6-7: Year-to-year annual variations (thin line) and five-year running average (thick line) of mean summer air temperature between May-September (T5-9) at UIB climate stations; Gilgit, Askole, Bunji, Astore, and Chilas. Note that temperature ranges on the y-axis differ between plots.**

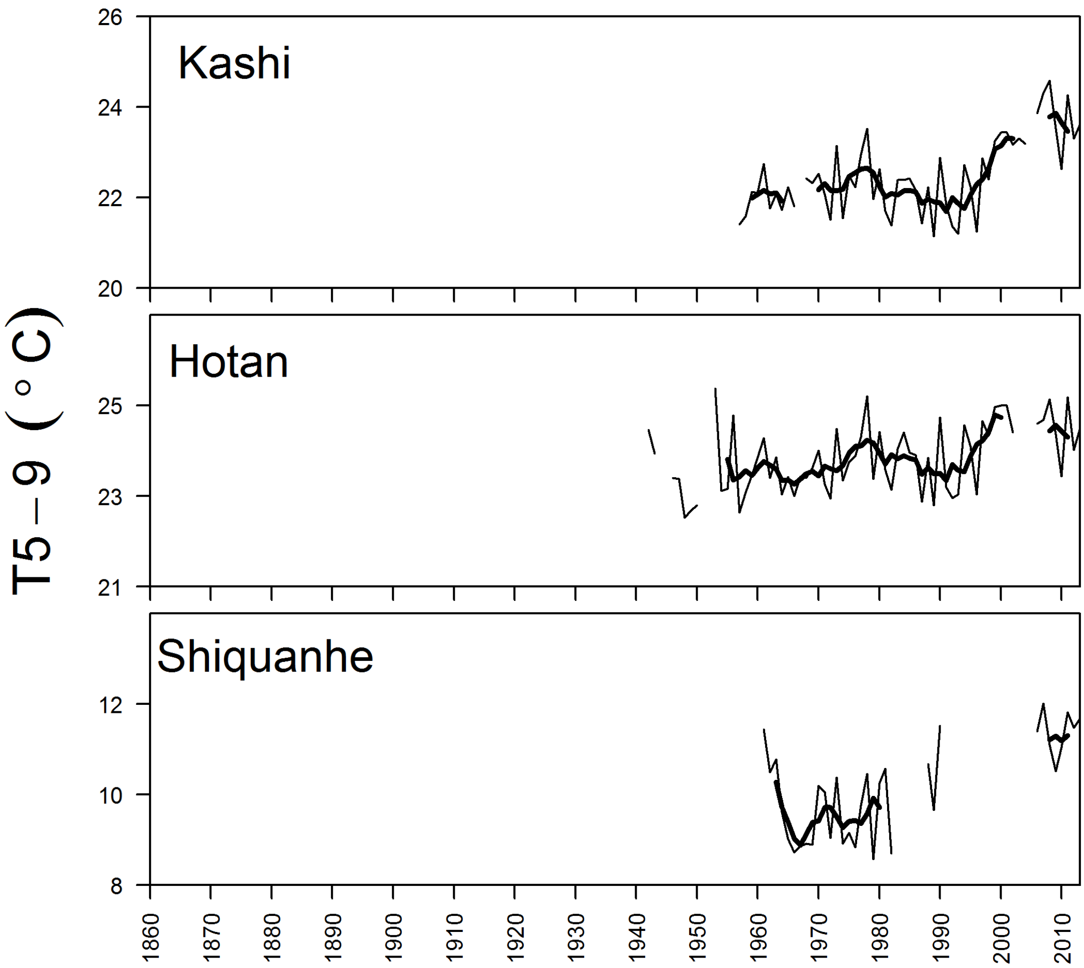


**Figure 6-8: Year-to-year annual variations (thin line) and five-year running average (thick line) of mean summer air temperature between May-September (T5-9) at Tibet climate stations; Kashi, Hotan and Shiquanhe. Note that temperature ranges on the y-axis differ between plots.**

three warmest years of the series in 1978, 1973 and 1971, before cooling to the 1980s and 1990s. Summer air temperatures increased to 2000s, but failed to reach maxima of the 1970s, albeit maxima quinquennia T5-9 was obtained in 2000 (22.14°C) (figure 6-9). Summer air temperature at Jhelum remained cool in the 1970s before steadily rising from the 1980s to a maximum in 2002 (31.96°C), declining slightly in the early 2000s. To the eastern end of the western Himalaya, long term mean summer temperatures generally fluctuated along an underlying falling trend. High summer temperatures recorded in the 1940s and 1950s at Sialkot and Amritsar have yet to be exceeded (Figure_6-9 and 6-10). At Sialkot, the time series of T5-9 was generally broken but declined from maxima in the 1940s to 2000s by around 1.2°C. Amritsar cooled to a minimum in the 1980s from which temperatures started to recover in the 1990s but remained relatively flat into the 2000s, failing to match the early maxima of the 1950s. Ludhiana followed a similar sequence, from its maxima in 1910s summer temperatures declined from the mid-1940s to a minimum in the 1960s/70s, before the record becomes unavailable. Shimla followed a distinctly dissimilar pattern to that of Ludhiana in the first half of its series; where summer temperature declined from its maximum in 1879 to a minimum in 1917. Temperatures stabilised in the 1920s and increased to the 1940s and 1960s, before cooling to the 1970s with some recovery in the 1980s, before the record ends prematurely in 1987. Mukteshwar showed similar evolutions, cooling to ~1910s from the early 1900 maxima, increasing from the 1920s to 1960s, before declining to cooler years in the 1970s, 1980s and 1990s, before rising to similar levels of the 1960s in the 2000s, with the warmest year of the record in 2012 (18.68°).

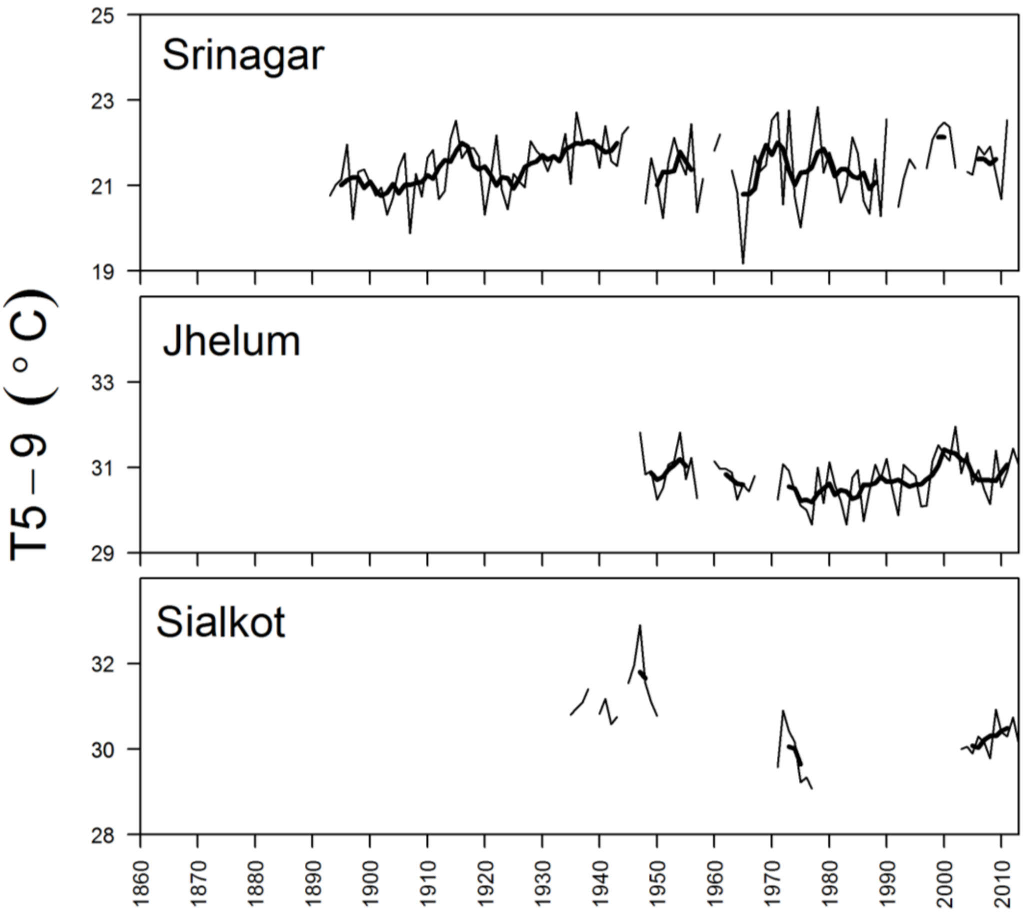


**Figure 6-9: Year-to-year annual variations (thin line) and five-year running average (thick line) of mean summer air temperature between May-September (T5-9) at western Himalayan climate stations; Srinagar, Jhelum, and Sialkot. Note that temperature ranges on the y-axis differ between plots.**

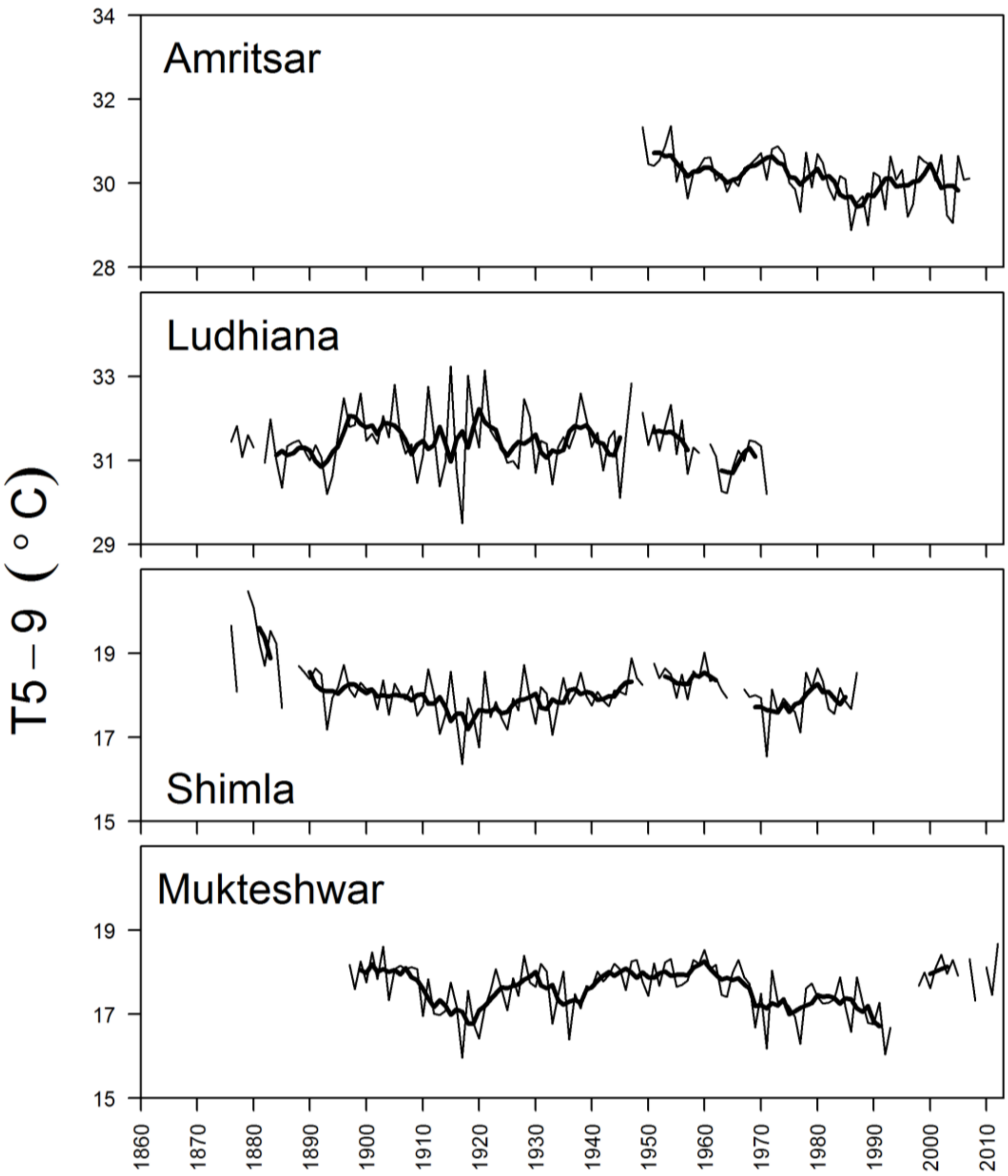


**Figure 6-10: Year-to-year annual variations (thin line) and five-year running average (thick line) of mean summer air temperature between May-September (T5-9) at western Himalayan climate stations; Amritsar, Ludhiana, Shimla, and Mukteshwar. Note that temperature ranges on the y-axis differ between plots.**

## 6.4. Long term year-to-year fluctuations of total annual discharge

### 6.4.1. UIB

Total annual discharge in the UIB was greatest in the glacierised Shyok and Hunza basins (average Q1-12 of 11.27 and 10.42 x10$^9$m$^3$), accounting for, on average, 14% and 13% of total annual discharge downstream at Tarbela dam. From year-to-year, annual variations of discharge at Shyok and Hunza varied considerably along an underlying declining trend from the 1970s to the late 1980s/early 1990s (figure 6-11). Annual total riverflow for the Hunza had no specific trend during the 1990's, with higher flows in 1990 and 1994 (both $11.92 x 10^9 m^3$) and lower flows in the intervening years, 1991, 1992 and 1993. Following low annual discharge in 1997 ($7.57 x 10^9 m^3$) (ranked 2nd lowest after 1989) riverflow slightly recovered in 1998 and 2001, but failed the reach levels of riverflow attained during the 1960's and 1970s, declining by around 24% from the 1960s/70s to early 2000s. At Shyok, riverflow recovered in the 1990s with the exception of 1993 which was the lowest recorded annual discharge of the record ($7.24 x 10^9 m^3$). Riverflow continued to rise into the 21st century, increasing by around 13% from the 1970s, with maximum quinquennium discharge centred on 2000. From the less glacierised Gilgit basin average total annual discharge was slightly less than that of Hunza and Shyok (average Q1-12 of $9.11 x 10^9 m^3$), contributing on average 12% of total annual discharge downstream at Tarbela. From the 1960s annual discharge reached a minimum in 1982 ($6.65 x 10^9 m^3$) before varying along a rising trend to the 1990s following the rising pattern of winter precipitation from UIB climate stations. Annual discharge remained relatively flat during the 1990s although was slightly above the long term average (1990 decadal average Q1-12 $9.37 x 10^9 m^3$), with higher flows in 1992, 1994, and 1998. Maximum quinquennium discharge was recorded in 2003 with the highest individual total in 2005. From the 1980s to 2000s riverflow increased by around 14%, but over the longer term, riverflow remained relatively stationary increasing by around 4% from the 1960s-2000s.

In the south facing nival catchment of Astore, riverflow at Doyien followed a similar pattern of riverflow at Gilgit; rising from the 1970s to 1990s with a brief decline 1982 and 1985. Unlike Gilgit, riverflow at Doyien inclined sharply in the 1990s to a maximum in 1996 ($6.27 x 10^9 m^3$). 1997 was the lowest recorded discharge of the record ($2.75 x 10^9 m^3$) with the 5 year running mean dropping to a minimum in 1999. Riverflow remained low in 2000 and 2001 (ranked 5th and 3rd lowest) before recovering in 2003, 2004 and 2005 to similar levels attained during the 1990s. Compared with Hunza, Shyok and Gilgit, average total annual discharge from the Astore catchment was considerably less (average Q1-12 of $4.3 x 10^9 m^3$) and contributed on average only 6% of total streamflow at Tarbela. Temporal variation of discharge on the Indus at Besham (average Q1-12 of $76.88 x 10^9 m^3$) mimicked the pattern of riverflow 135km downstream at Tarbela (average Q1-12 of $78.25 x 10^9 m^3$)(figure 6-11). Annual total riverflow at Besham and Tarbela remained relatively stable in the 1960s and 1970s but varied

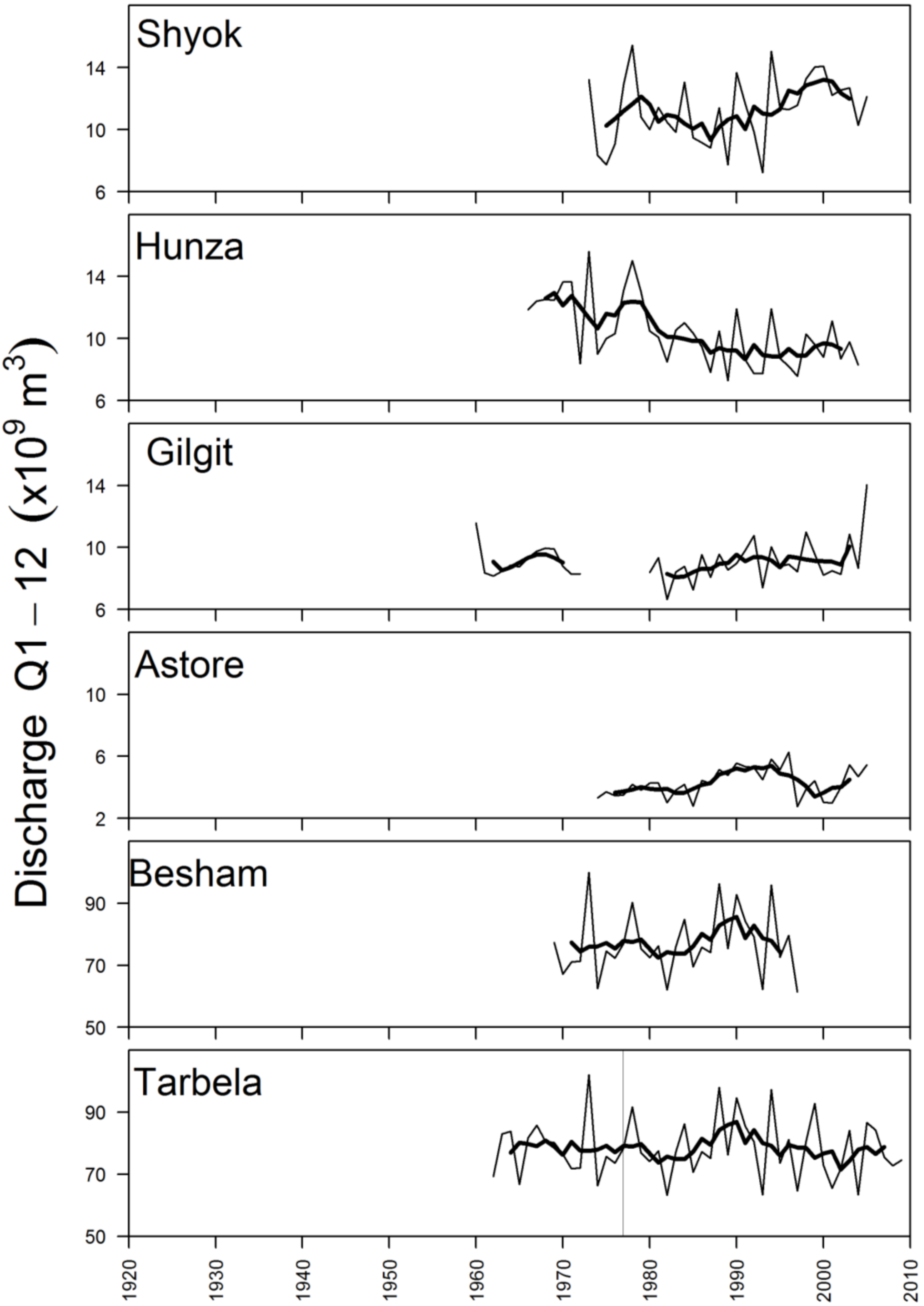


**Figure 6-11: Year-to-year variations (thin line) and five-year running average (thick line) of total annual discharge for calendar years, January-December (Q1-12), at UIB gauging stations. Note discharge values on the y-axis differ. Vertical line indicates the completion of Tarbela dam in 1977.**

considerably from year-to-year. A slight downward trend affected the early 1980s, reaching a minimum in 1982 (ranked 1st lowest at Tarbela and 2nd lowest at Besham), but increased thereafter to maxima of both series in the late 1980s/ early 1990s with quinquennium maxima centred on 1990, despite reducing riverflows from its tributaries Shyok and Hunza. Riverflow declined slightly in the mid-1990s but remained above the series average (1990s decadal average Q1-12 of 78.46x10$^9$m$^3$ at Besham and 81.40x10$^9$m$^3$ at Tarbela) with high flows in 1994 and 1999 and low flows in 1993 and 1997. From 2000 riverflow at Tarbela remained subdued with the decade average failing to reach the series average and the 5 year running mean dropping to a minimum in 2002, nonetheless, long term riverflow at Tarbela had no specific trend, declining by around 4% from the 1960s-2000s.

### 6.4.2. Western Himalayas

Prior to the construction of Mangla dam, annual riverflow at Mangla declined by around 10% from the 1920s to 1940s to a minimum in 1946 (24.59x10$^9$m$^3$), before increasing again by around 14% to the 1950s attaining the series maximum in 1959 (42.8x10$^9$m$^3$). Following the construction of Mangla dam between 1962-1967 annual riverflow declined to the series minimum in 1971 (15.56x10$^9$m$^3$) but varied along an increasing trend from 1972, mirroring the pattern of winter precipitation and following similar variations of riverflow to the Astore basin north of the Jhelum catchment. With progressively higher flows from 1973 to 1996 riverflow inclined to maximum quinquennium in 1993, rising by about 24% from the 1970s-1990s, but fell sharply in 1997 to the second lowest annual total of the series in 2000 (figure 6-12) (16.38x10$^9$m$^3$). Annual total discharge at Marala (average Q1-12 of 32.54x10$^9$m$^3$) followed a broadly similar pattern of riverflow to that of Mangla; declining from the series maximum in 1959 (46.57x10$^9$m$^3$) to a period of low flows in the 1960s and 1970s, increasing to the 1980s, until remaining relatively flat in the 1980s and 1990s. Riverflow decreased considerably in 1999, 2000, 2001, and 2004, all failing to reach the series average, and only just exceeding the series average in the intervening years 2003, 2005 and 2006.

In the upper Sutlej at Khab (average Q1-12 of 6.87x10$^9$m$^3$), annual discharge declined by around 31% from the 1970s to 1990s/2000s to a minimum in 2001 (2.49x10$^9$m$^3$) (figure 6-13). Natural riverflows downstream at Rampur, Luhri and Suni followed similar year-to-year variations, declining to a series minimum in 2004 and 2001, respectively, with riverflows declining by around 15% between 1970 and 1990/2000.

Downstream, in the lower Sutlej, riverflow at Bhakra was augmented by waters diverted from the Beas-Sutlej link canal after 1977. Discharge at Bhakra (average Q1-12 of $15.49 \times 10^9 m^3$) declined by around 12% from the 1920s-1930s but increased by around 18% from the 1930s-1950s to the series maximum in 1955 ($23.78 \times 10^9 m^3$). From the 1950s discharge declined to the 1970s, declining by around 5% from the 1970s-1990s/2000s. Runoff is enhanced considerably downstream in the Sutlej, where the volume of discharge increased on average by 126% from its headwaters at Khab in the arid region of Tibet to the dam at Bhakra, and by 62% from Khab to Rampur, thus highlighting the influence of precipitation on the riverflow downstream.

## 6.5. Year-to-year variability of total annual discharge

Year-to-year annual variability of runoff, expressed using the coefficient of variation, for a 21-year period (1980-2000) for which data was available for all gauging stations across the study area showed no specific trend between CV of annual runoff and the location of a gauging station across a west-east transect (table 6-1). The highest calculated year-to-year variation of runoff was observed at Khab (CV= 0.24) on the Sutlej where its headwaters originate from melting of snow and ice in arid Tibet. High variability was also shared with the relatively low ice covered catchments; Jhelum at Mangla and the Astore at Doyien (CV=0.22). Relatively high runoff variability was also measured in the glacierised Shyok (CV=0.19) and Hunza (CV=0.15) basins, both basins having lower variability than less ice covered basins. Coefficient of variation of annual runoff reduced downstream on the Sutlej, CV reducing by half from Khab to Rampur (CV= 0.12), and reducing further to (CV=0.11) at Suni and Bhakra from the persistent monsoon precipitation which reduces northward, rarely penetrating the large Tibetan proportion of the Sutlej basin. Intermediately glacierised basins; Gilgit, Tarbela, and Marala shared relatively low year-to-year variations of runoff similar to that in the Sutlej, although icemelt provides the moderating influence on runoff, rather than the monsoon in the east, as in warm dry years icemelt compensates for the reduced precipitation over the ice free area of a catchment. Surprisingly, the precipitation driven Sutlej basin at Bhakra (CV=0.11) had the lowest annual variability of any major reservoir gauging stations between 1980 and 2000, being marginally lower than moderately glacierised Tarbela, and Marala (CV=0.13). Over a period of 39 years (1962-2000) for which data was available for all four major reservoirs, low runoff variability was maintained at Tarbela and Bhakra (CV=0.12) but increased at Marala (CV= 0.16), Mangla still maintaining the highest variation of riverflow at CV= 0.22. However, over a period of 43 years (1962-2004), minimum variability was obtained at Tarbela (CV=

0.12), CV slightly increasing at Bhakra to 0.14. Minimum variability of annual runoff obtained at Tarbela reflects the compensation effect of runoff in a mountain catchment where runoff from the glacierised sub basins (Shyok and Hunza) have opposing runoff responses to nival catchments, which are non-synchronous in yearly variation.

Longer records are preferred and reflect the true variability of riverflow in a catchment, as shorter records may be biased toward a period of sustained climatic variation which influences the calculated variability. In the western Himalayas, Longer periods of discharge records tended to yield higher values of CV, except for Mangla which had the same CV for a 21 and 39 year period (table 6-1), whilst for gauging stations in the UIB, shorter records yielded higher values of runoff variability, CV declining with longer periods of records.

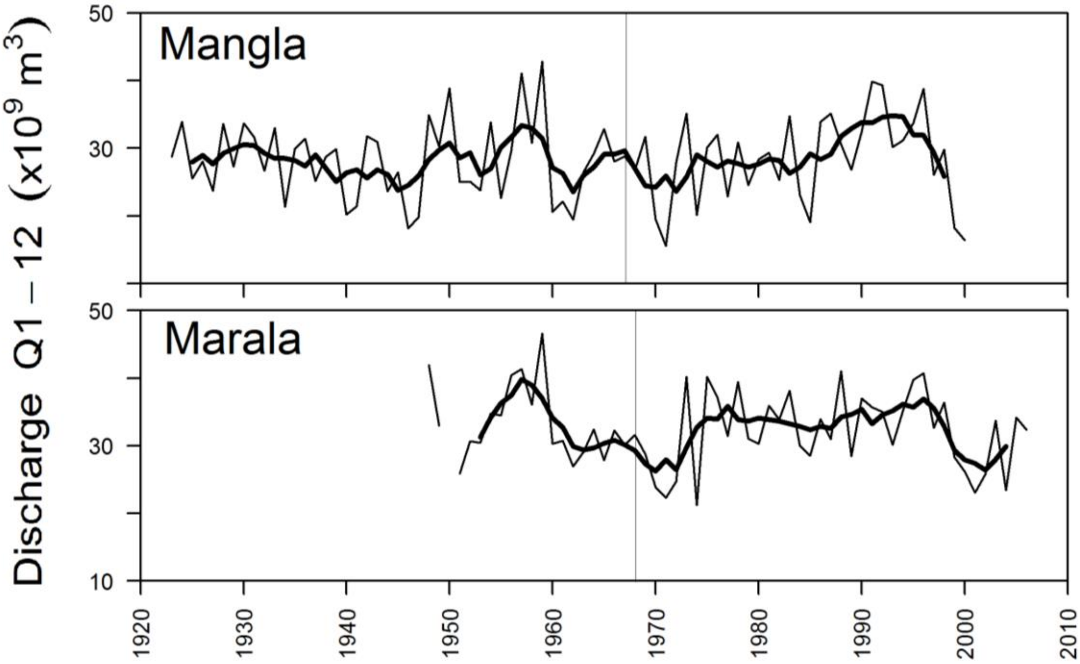


**Figure 6-12: Year-to-year variations (thin line) and five-year running average (thick line) of total annual discharge for calendar years, January-December (Q1-12), at western Himalayan gauging stations. Vertical line indicates the completion date of respective dams; 1967 at Mangla and 1968 at Marala.**

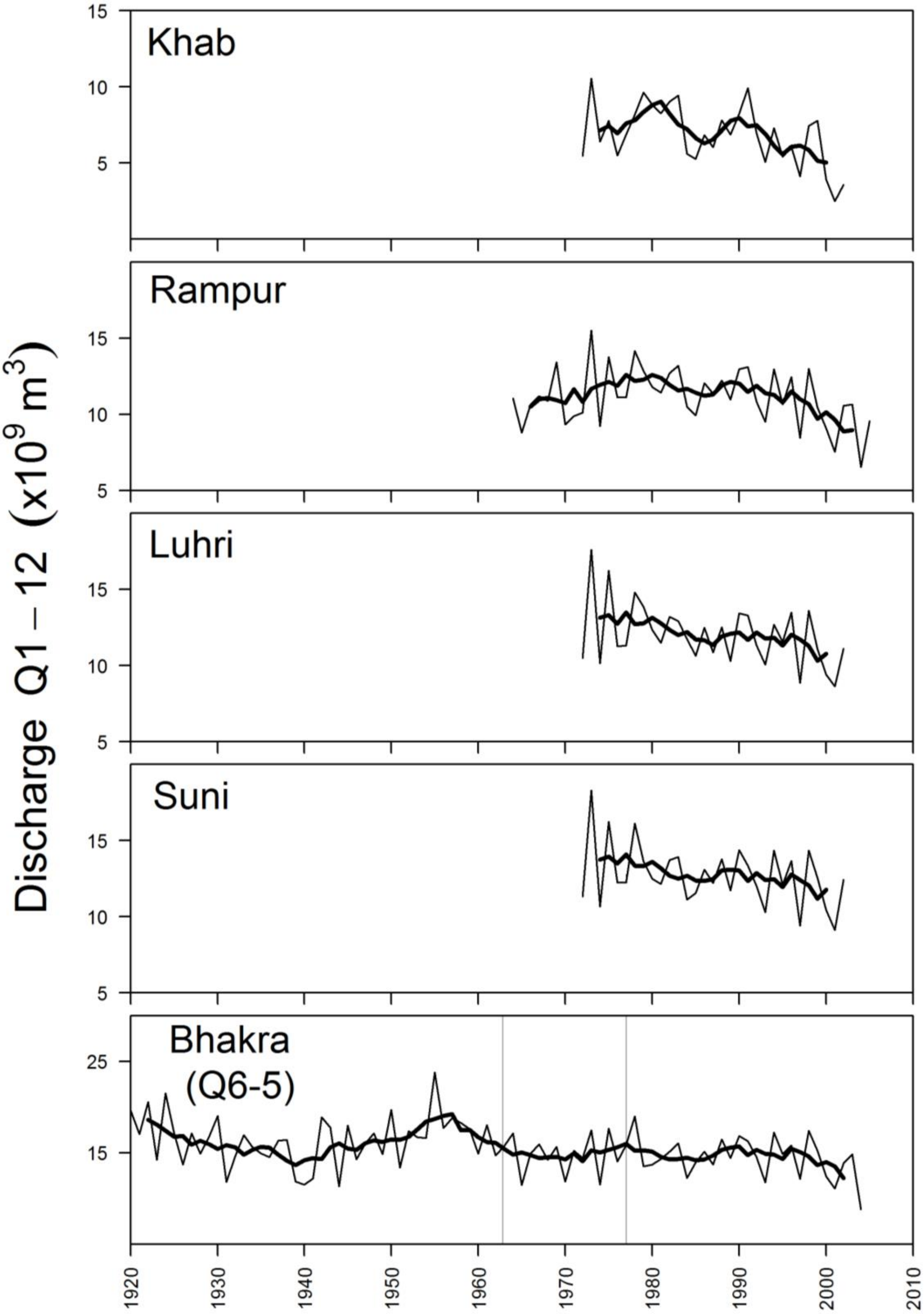


**Figure 6-13: Year-to-year variations (thin line) and five-year running average (thick line) of total annual discharge for calendar years, January-December (Q1-12), at gauging station on the Sutlej. Note however that Bhakra discharge is from June-May (Q6-5). Vertical lines indicate the construction of Bhakra in 1963 and the Beas-Sutlej link in 1977.**

### 6.5.1. UIB

During the 25-year period (1980-2004) for which runoff data are available for all five gauging stations in the Indus catchment, variability of annual runoff was greatest in the least ice covered Astore basin (CV= 0.22), least in the intermediate Gilgit basin (CV= 0.12), but then increased with increasing glacierisation to CV= 0.14 in the Hunza and CV= 0.17 in the Shyok, creating a typical asymmetrical U-shaped relationship between CV and ice area of a basin (figure 6-14). The relationship between ice area and CV exhibits a negative correlation (r= -0.11), with variability of runoff being less in the glacial Shyok catchment by comparison with that the nival Astore catchment, reflecting the moderation of riverflow by the large ice area. The relationship between CV and percentage glacierisation of a basin is unorthodox, creating an asymmetrical n-shape relationship between percentage glacierisation and CV, contrary to the typical U-shape relationship (as discussed in chapter 4) (figure 6-14).

## 6.6. Year-to-year variability of precipitation variables

Coefficient of variation for totals of precipitation between June-September and November-May for periods of years which are equivalent to the periods of runoff variation measurements are presented in table's 6-2 and 6-3. Variations in winter and summer precipitation were generally higher than variations of total annual runoff, with the exception of P6-9 at Dehradun district between 1980 and 2000, which are lower than discharge variations at Khab, Gilgit and Mangla. Intra-annual variations of the summer monsoon were lower in the east where precipitation totals were higher, CV of P6-9 gradually increasing westerly as precipitation totals reduced. For example; between 1962-2000 variations of summer precipitation were lowest at Dehradun district (CV= 0.27, mean P6-9 of 1748.1mm) and Shimla district (CV= 0.28, mean P6-9 of 797.2mm), increasing slightly to CV=0.38 at Srinagar district (mean P6-9 of 237.3mm), before increasing to CV= 0.46 at Astore on the border of the western Himalaya (mean P6-9 of 96.4mm), to the highest variability at Chilas in the UIB (CV= 0.74, mean P6-6 of 42.3mm).

Coefficients of variation of winter precipitation were generally higher than those variations of summer precipitation, except at Astore, Chilas and Srinagar district, where variations of winter precipitation, for the respective periods, had lower year-to-year variability. Over 39 years (1962-2000) variability of winter precipitation had no specific trend from west to east. Year-to-year variations of winter precipitation were lowest at Srinagar district (CV = 0.29) and Astore (CV= 0.34), where westerly disturbances dominate at the border of the western Himalaya and Karakoram. Variability increased toward the east at Shimla district (CV= 0.44) and Dehradun district (CV= 0.53) with weakening winter westerly disturbances, but surprisingly intra-annual variations of P11-5 were highest in the UIB at Chilas (CV= 0.69) and Gilgit (CV= 0.60), in spite of the heavy influences of westerly disturbances in the UIB, but actually reflects the

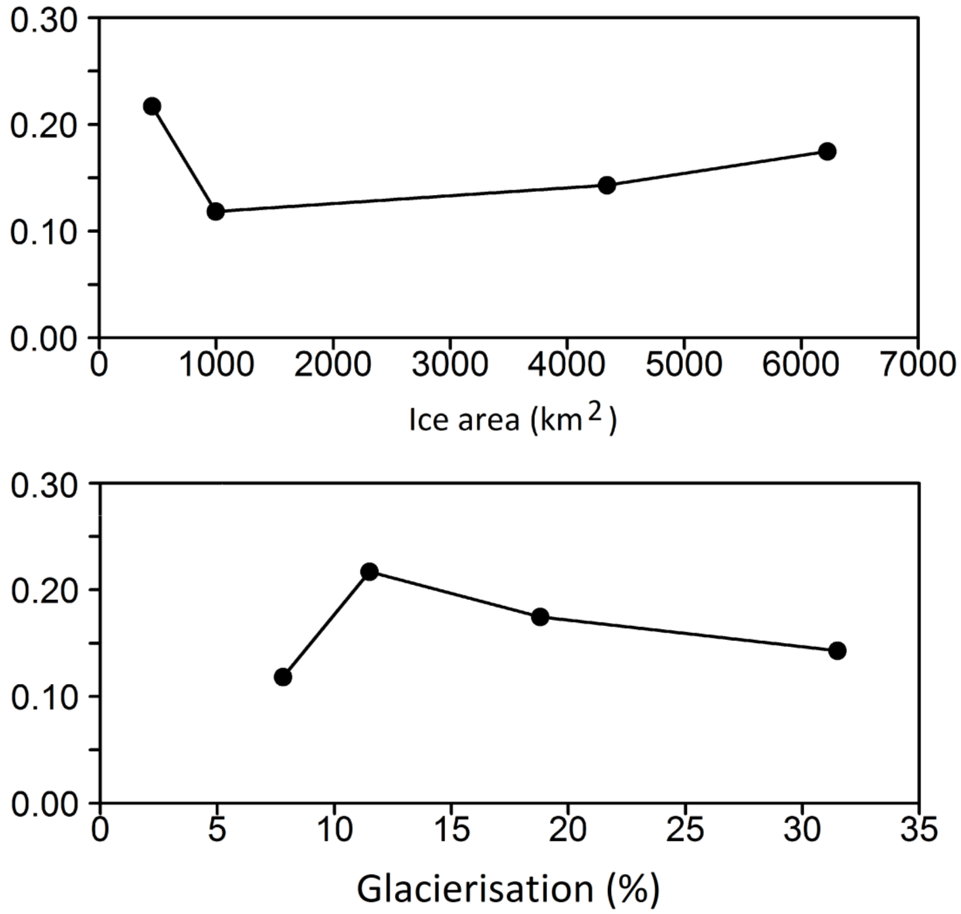


**Figure 6-14: Coefficient of variation (CV) of annual total runoff (Q1-12) from the four study basins in the UIB for the period 1980-2004 against ice area of a basin.**

low lying meteorological stations which have the lowest annual totals of precipitation over the entire study area.

Intra-annual variations of total annual precipitation from January to December (P1-12), which includes both summer and winter totals of precipitation, were considerably less than variations of summer and winter precipitation at all climate stations (table 6-4), as low precipitation totals in one season may be offset by large increases of precipitation in another season, which tends to moderate annual variations.  Variability of total annual precipitation were generally lower toward the east at Dehradun district, Shimla district, and Srinagar district, where annual totals of precipitation were greater, variability increasing at Chilas, Bunji, and Gilgit, where annual totals of precipitation were typically less than ~200mm$^{yr-1}$.

**Table 6-1: Coefficient of variation of total annual discharge from basins in the UIB and Himalaya over periods between 21 and 43 years.**

| Period | Length of period (years) | Basin Tarbela | Mangla | Marala | Bhakra | | Shyok | Astore | Gilgit | Hunza | | Khab | Rampur | Luhri | Suni |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1980-2000 | 21 | 0.13 | 0.22 | 0.13 | 0.11 | | 0.19 | 0.22 | 0.12 | 0.15 | | 0.24 | 0.12 | 0.12 | 0.11 |
| 1980-2004 | 25 | 0.13 | | | | | 0.17 | 0.22 | 0.12 | 0.14 | | | | | |
| 1972-2000 | 29 | | | | 0.13 | | | | | | | 0.24 | 0.14 | 0.16 | 0.15 |
| 1962-2000 | 39 | 0.12 | 0.22 | 0.16 | 0.12 | | | | | | | | | | |
| 1962-2004 | 43 | 0.12 | | 0.17 | 0.14 | | | | | | | | | | |

**Table 6-2: Coefficient of variation of total summer precipitation (P6-9) from stations in the UIB and Himalaya over periods between 21 and 39 years.**

| Period | Length of period (years) | Station Gilgit | Skardu | Bunji | Astore | Chilas | | Srinagar | Srinagar district | Shimla district | Dehradun district |
|---|---|---|---|---|---|---|---|---|---|---|---|
| 1980-2000 | 21 | 0.44 | | 0.51 | 0.43 | 0.61 | | 0.35 | 0.40 | 0.27 | 0.21 |
| 1980-2004 | 25 | 0.41 | | 0.50 | 0.40 | 0.62 | | | | | |
| 1972-2000 | 29 | 0.40 | | 0.49 | 0.43 | 0.64 | | | | 0.28 | 0.27 |
| 1962-2000 | 39 | 0.45 | | 0.54 | 0.46 | 0.74 | | | 0.38 | 0.28 | 0.27 |

**Table 6-3: Coefficient of variation of total winter precipitation (P11-5) from stations in the UIB and Himalaya over periods between 21 and 39 years.**

| Period | Length of period (years) | Station Gilgit | Skardu | Bunji | Astore | Chilas | | Srinagar | Srinagar district | Shimla district | Dehradun district |
|---|---|---|---|---|---|---|---|---|---|---|---|
| 1980-2000 | 21 | 0.61 | 0.41 | 0.56 | 0.33 | 0.60 | | | 0.28 | 0.42 | 0.48 |
| 1980-2004 | 25 | 0.61 | | 0.59 | 0.34 | 0.60 | | | | | |
| 1972-2000 | 29 | 0.65 | | 0.52 | 0.34 | 0.69 | | | | 0.48 | 0.51 |
| 1962-2000 | 39 | 0.60 | 0.43 | 0.58 | 0.34 | 0.69 | | | 0.29 | 0.44 | 0.53 |

**Table 6-4: Coefficient of variation of total annual precipitation (P1-12) from stations in the UIB and Himalaya over periods between 21 and 39 years.**

| Period | Length of period (years) | Station Gilgit | Skardu | Bunji | Astore | Chilas | | Srinagar | Srinagar district | Shimla district | Dehradun district |
|---|---|---|---|---|---|---|---|---|---|---|---|
| 1980-2000 | 21 | 0.36 | | 0.37 | 0.27 | 0.46 | | | 0.23 | 0.17 | 0.16 |
| 1980-2004 | 25 | 0.36 | | 0.41 | 0.26 | 0.47 | | | | | |
| 1972-2000 | 29 | 0.37 | | 0.36 | 0.26 | 0.54 | | | | 0.19 | 0.22 |
| 1962-2000 | 39 | 0.37 | | 0.36 | 0.26 | 0.54 | | | | 0.19 | 0.22 |

# 7. Discussion

## 7.1. Temporal variations of air temperature, precipitation and riverflow

River flow from partially glacierised basins arises from seasonal melting of snow cover, summer precipitation and melting of glacier ice in varying proportions and differing absolute quantities along west –east axes of the Karakoram and western Himalaya. In the UIB and Karakoram winter precipitation nourishes glaciers in the form of snow at high altitudes where precipitation increases with elevation despite low quantities of observed total precipitation from valley stations (table 5-1) (e.g. Immerzeel et al 2012). Flows from highly glacierised basins generally follow fluctuations of air temperature which affect melting of glaciers, while less glacierised basins mimic temporal variations of precipitation, predominantly winter precipitation in western catchments. Eastern catchments which are less glacierised follow patterns of precipitation, mainly the summer monsoon, which accounts for almost ¾ of total annual precipitation. Meteorological inputs ultimately determine riverflow outputs. Total river flow in a glacierised basin may decrease by the following climatic interactions;

1) Decreased summer temperatures reducing glacier melt water production
2) Increased winter precipitation over the glacier surface which limits the duration of clean ice exposed to melting during the ablation season
3) Increased summer precipitation falling as snow on the glacier protects the underlying ice from ablation by increasing surface albedo
4) A period of sustained climatic warming causing the deglaciation discharge dividend, the component of runoff derived from net melting of ice, to decline following a significant reduction of ice area.

Discharge from less glacierised basins may decline during a period of reduced precipitation, winter precipitation dominating riverflow in the west and summer monsoon precipitation in middle and lower reaches of catchments in the east.

### 7.1.1. UIB and Karakoram

Runoff records for sub-basins in the UIB are relatively short, with earliest records starting in 1960, by comparison to time series of meteorological records which commence from 1892. Plots of discharge from partially glacierised basins are compared with glaciological relevant variables of P6-9, P11-5 and T5-9 from nearby meteorological stations (Figures 7-1 – 7-4). Since the commencement of discharge records at Dainyor Bridge in 1966 riverflow from Hunza declined by around 24% from 1966/75 to 1995/04, following a reduction of 0.26°C and 1.9°C of mean May to September temperatures at Gilgit and Bunji for the respective period (figure 7-1). Mean summer temperatures at Gilgit and Bunji fluctuated on a declining trend from the 1970s, declining to a minimum in the late 1980s/early 1990s and remaining down in the 2000s. For years where runoff records coincide with air temperature records, 1973, 1977 and 1990 stand out as warm years at Gilgit, with 1971 and 1973 standing out as warm years at Bunji. Warm years in the 1970s enhanced runoff totals at Hunza, 1973 and 1978 being ranked 1st and 2nd highest runoff totals, respectively. Low annual total runoff in 1972, 1974 and 1975 resulted from reduced mean summer air temperatures. Runoff reduced to a minimum in 1989 where temperatures at Gilgit and Bunji also declined to a series minimum. Reduced winter precipitation at Gilgit in the 1980s exposed additional ice to melting during the summer melt season as the transient snowline raised sooner, but annual runoff at Hunza remained subdued following reduced summer air temperatures and increasing summer precipitation. Summer precipitation falling as snow on the ablation area of the glacier can reduce melting of ice and thus runoff by raising the albedo (e.g. Collins 1982). Total annual summer precipitation increased by ~45% at Gilgit and Bunji for the common period of record with runoff, 1966/75 to 1995/04. Wet summers at Gilgit occurred in 1978, 1988, 1989 and 1997, with wet years at Bunji in 1986, 1993, 1997, 2000 and 2003, annual runoff at Hunza being reduced in all years except for 1978 where summer air temperatures were particularly warm (figure 7-1).

The relationship between winter precipitation and runoff was complex. Wet winters at Gilgit in 1973 and 1979 coincided with warmer summers and high specific runoff. Rapid depletion of the winter snow pack during spring allowed the transient snowline to ascend sooner, thus allowing sufficient melting and runoff from bare ice. In winters with above average snow cover, but with cooler summer temperatures (1996, 1998, 1999, 2003), runoff was reduced as the transient snowline in summer ascended slower and limited the total exposure and duration of underlying ice to melting. Glaciers in the Hunza basin have observed ice thickening around the upper and lower glacier (Hewitt 2005). Summer cooling since the 1970s and increasing seasonal precipitation since the

1990s have reduced riverflows in the Hunza basin, thus agreeing with the so called Karakoram anomaly.

Specific runoff from Hunza (0.756m) is lower than in observed glacierised basins in the Swiss Alps and the North American Cascades (Collins 2008, Meier and Tangborn 1961). Lower, ice free areas of valleys in the Karakoram-Hunza are arid. The winter snowline in the UIB descends to elevations ~2300m a.s.l (Immerzeel et al 2009) and thus do not receive enough precipitation over its entire basin area to compensate for reduced icemelt in cooler summers. Q1-12 at Yugo in the Shyok catchment mimicked the pattern of discharge at Hunza; declining from the late 1970s to the late 1980s/ early 1990s. However, from the 1990s riverflow increased to the early 2000s, rising by around 13% from the first to last decade of discharge measurements, 1973/82-1996/05. The pattern of riverflow at Shyok reflected year-to-year variations of T5-9 in the Western Himalaya and Tibet, where temperatures increased by around 0.4°C and 0.6°C at Srinagar and Hotan between 1973/82-1996/05 (figure 7-2). Mean summer air temperature declined from the late 1970s to the late 1980s before recovering in the 1990s and 2000s. A discharge maximum in 1978 was reflected by the warmest summer at Srinagar and Hotan where records overlap. High specific discharge in 1973, 1990, and 1994 was also reflected by above average summer air temperatures which enhanced melting of ice. Reduced runoff in 1975, 1989 and 1973 reflected reduced energy availability to melt snow and ice.

Summer and winter precipitation at Skardu had a weak relationship with discharge at Shyok despite precipitation increasing by ~7% and 25%, respectively, for the period 1973/82-1990/99. P6-9 at Skardu was down in the early 1980s but varied greatly from year-to-year increasing to the 1990s. P11-5 at Skardu was high in the early 1970s; 1973, 1974 and 1975, but fell in 1976, increasing on an underlying trend to the 1990s with maxima in 1996. Precipitation totals are reduced and are highly variable from year-to-year in the eastern Shyok basin, where a large proportion of the basin lies in Ladakh and Tibet, and much of the area remains less than 20% snow covered during winter and spring (Immerzeel et al 2009). Aridity of the basin is expressed with 0.341m specific runoff. Smaller snow cover area allows an extended melt season as the duration of clean ice exposed to melting is longer. Increased discharge from Shyok and increased summer air temperatures from the western Himalaya and Tibet since the 1990s is supported by a loss of glacier area toward the east where small glaciers of the Kang Yatze massif below the Indus River at Leh declined by 14.3% from 1962-2010 (Schmidt & Nusser 2012). Decreases of glacier area of Siachen glacier in the northwest Shyok

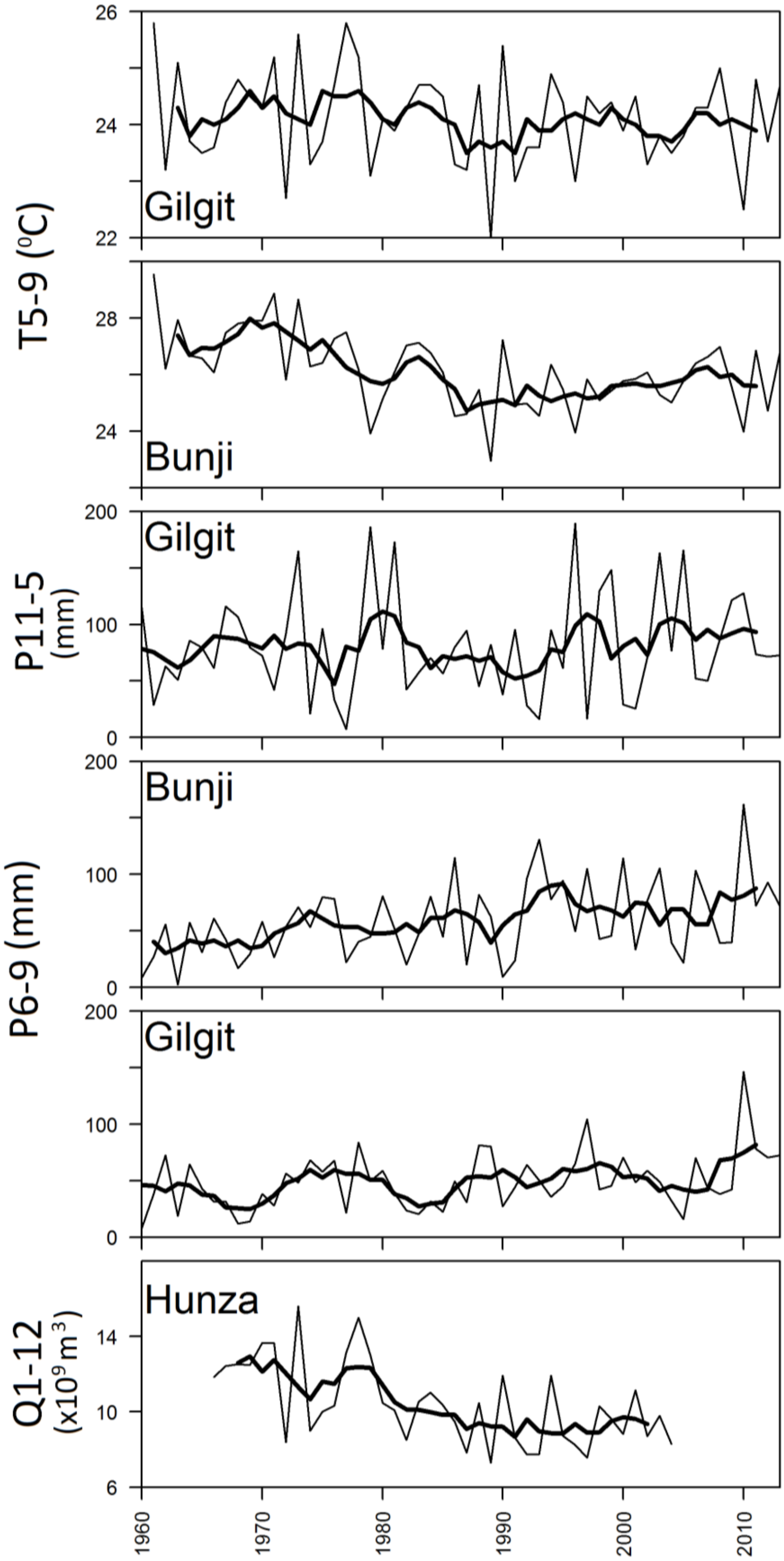


**Figure 7-1: Year-to-year variations of mean summer air temperature at Gilgit and Bunji, totals of winter precipitation at Gilgit, totals of summer precipitation at Bunji and Gilgit, and annual totals of discharge Dainyor bridge, Hunza from 1960-2013.**

basin have also been reported from 1986/87 and 1990/91 (Bhutiyani et al 1999) and between 1989 and 2006 (Rasul et al 2008), which under a period of climatic warming would augment riverflow. Contrary to these findings a stable mass balance of Siachen glacier was inferred from ice surface velocities between 2001/02-2009/10 (Heid and Kaab 2012), and a study area covering 2499km$^2$ of ice in the Shyok basin reported heterogeneous glacier response but an overall slight area change of +0.2% between 1973-2011 (Bhambri et al 2012). Glacier area changes might be attributed to surging glaciers which are common among Karakoram glaciers (Bhambri et al 2012). Increasing temperatures since the 1990s enhanced melting of glacier ice and runoff. Although glaciers may not have down wasted significantly, smaller glaciers in the east have responded quicker to climate change (Schmidt and Nusser 2012), whereas, larger glaciers in the western Shyok basin may be in-line with the Karakoram anomaly where increasing precipitation and declining summer air temperatures may have neutralised glacier mass balance.

Specific runoff at Astore (1.104m) is similar to ice free/ near ice free catchments in Switzerland (Collins 2008). Runoff in the least ice covered basin is augmented by plentiful precipitation captured by topography of the Nanga Parbat massif, precipitation in the Astore basin reaching lower altitudes than Karakoram basins. Discharge at Astore increased by around 14% from the first to the last decade of record, 1974/83-1996/05, mimicking the pattern of P11-5 at Astore which increased by 13.5% for the same period (figure 7-3). The wettest winter in 1996 was reflected by maxima discharge. Minimum discharge in 1997, which was over two times less than 1996, reflected reduced precipitation. P6-9 at Astore increased by 13% between 1974/83 and 1996/05, wet summers generally having a positive influence on runoff. Summer precipitation declined slightly after the 1990 maxima. Discharge in the moderately glacierised Gilgit basin also reflected patterns of P11-5 from its neighbouring meteorological station, Gilgit, where P11-5 increased by 29% and discharge by 4% between 1960/69 and 1996/05 (figure 7-4). Winter precipitation over the Karakoram is expected to increase up to the year 2100 (Ridley et al 2013) which would continue to replenish streamflow in basins with reduced ice cover. Riverflow declined to the 1980s but increased to the 1990s and 2000s with maxima in 2005. T5-9 at Gilgit had a weak relationship with discharge, summer temperatures declining from the 1970s to 1990s. Summer precipitation at Gilgit generally had a negative response on runoff. Wet winters followed by dry summer's yielded high quantities of runoff, most notable in 1960 and 2005.

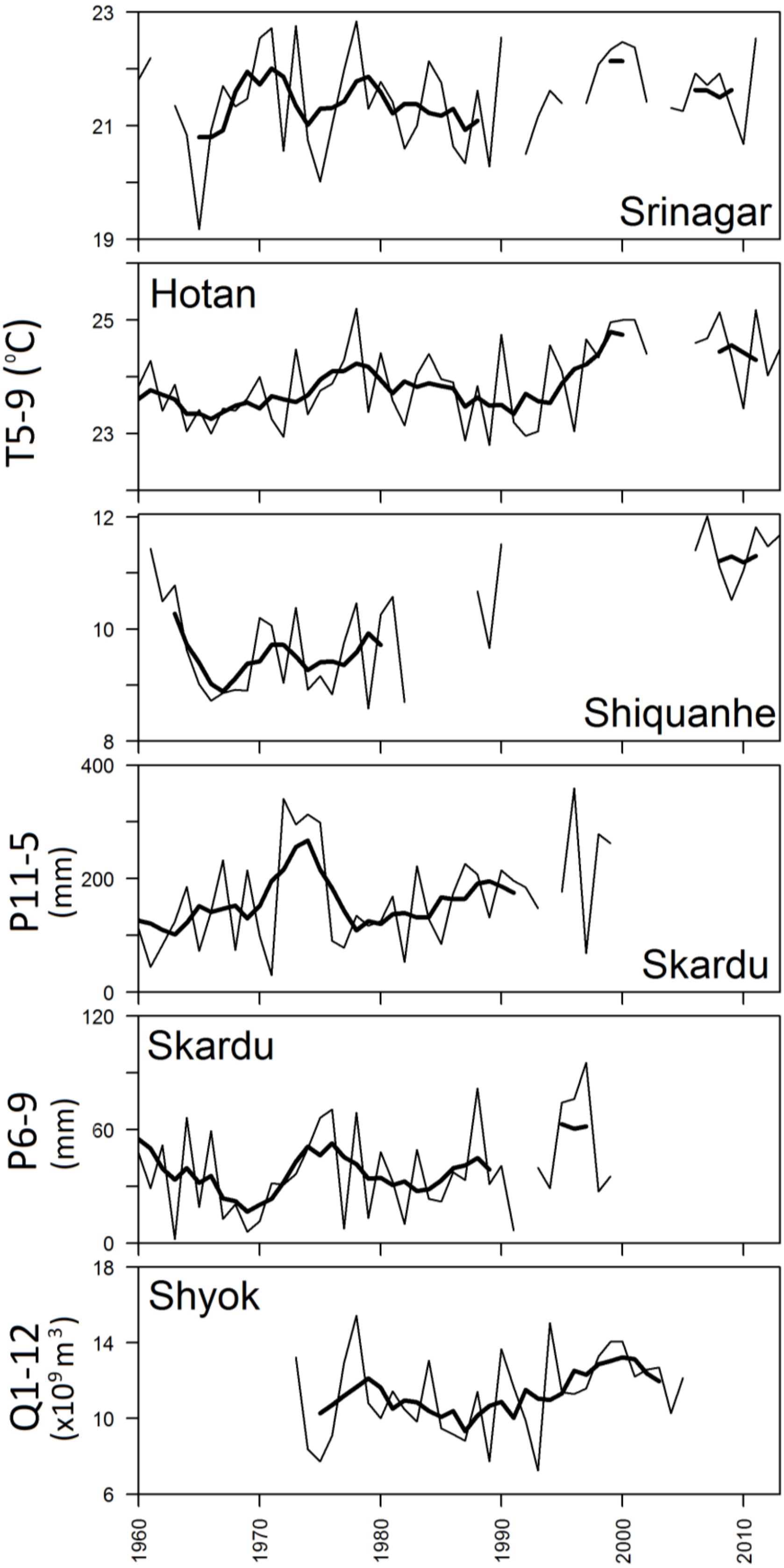


**Figure 7-2: Year-to-year variations of mean summer air temperature at Srinagar, Hotan and Shiquanhe, totals of winter and summer precipitation at Skardu, and annual totals of discharge at Yugo, Shyok from 1960-2013.**

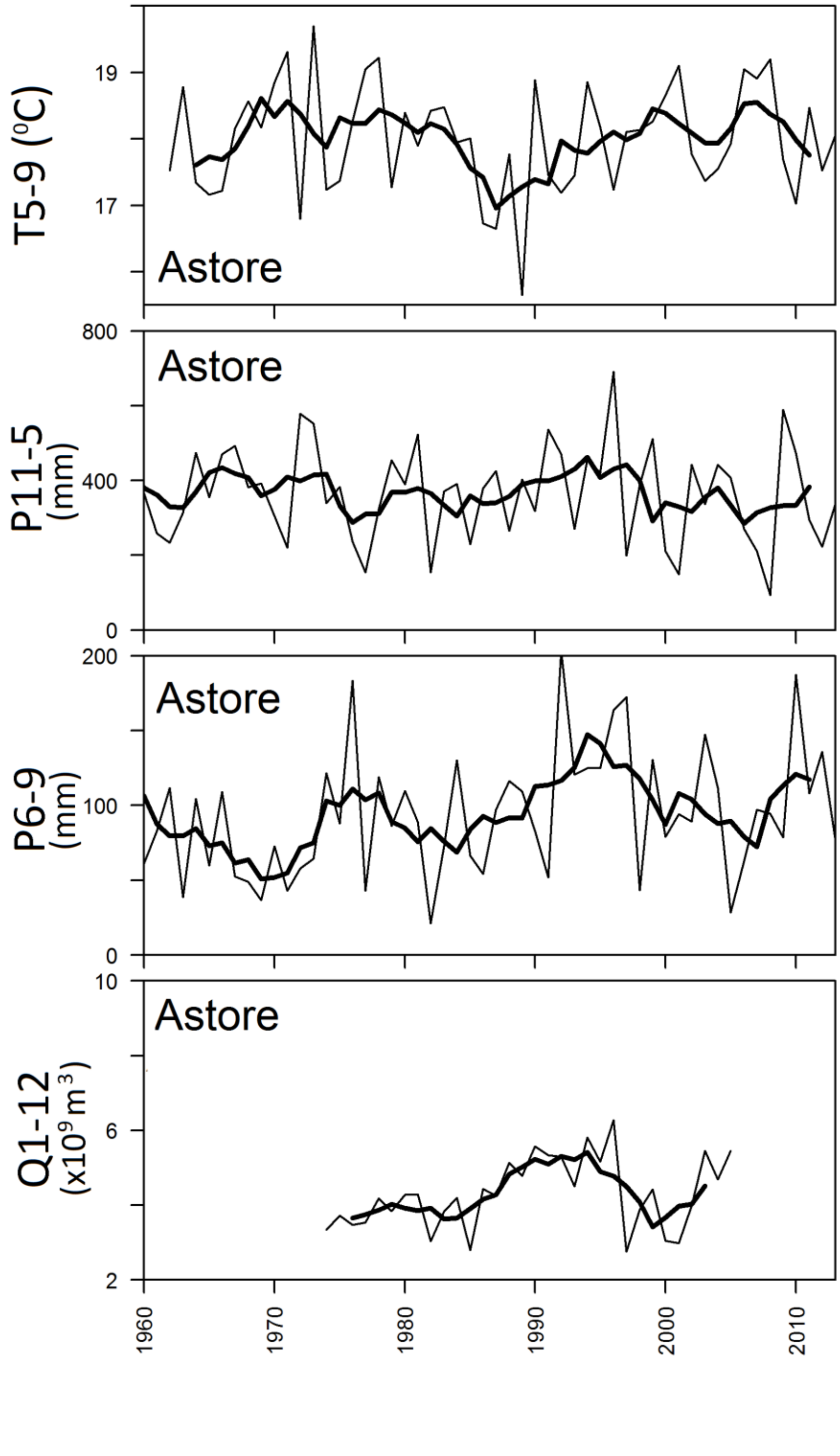


**Figure 7-3: Year-to-year variations of mean summer air temperature at Astore, totals of winter and summer precipitation at Astore, and annual totals of discharge at Doyien, Astore from 1960-2013.**

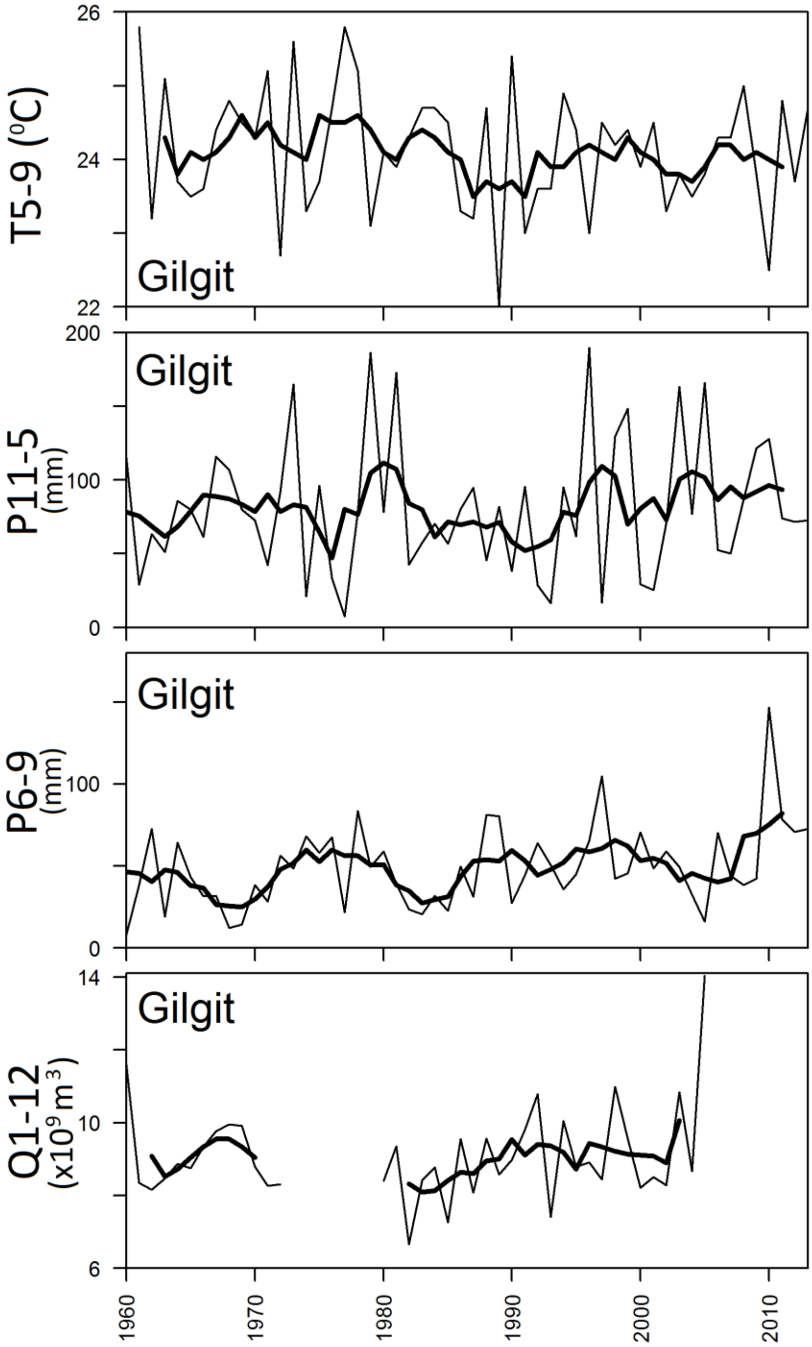


**Figure 7-4: Year-to-year variations of mean summer air temperature at Gilgit, totals of winter and summer precipitation at Gilgit, and annual totals of discharge at Gilgit from 1960-2013.**

### 7.1.2. Western Himalaya - Sutlej

Specific runoff in the upper Sutlej at Khab (0.166m) was the least of any gauging station in the whole study area which reflects aridity of the basin where around 60% of the basin area above Bhakra lies in the arid region of Tibet (calculated from Singh and Jain 2002). Riverflow at Khab, which accounts for on average 44% of the riverflow at Bhakra, is fed by melting of small glaciers and winter precipitation. Discharge at Khab in the low glacierised Sutlej basin is unrelated to summer air temperature at Mukteshwar. T5-9 at Mukteshwar decreased from the 1960s to the cool 1970s and 1980s, rising from a minimum in 1992 to similar levels of the 1960s in the 2000s (figure 7-5). Annual total discharge at Khab declined from the 1970s by around 31% between 1972/81 and 1993/02, mimicking an 11% decrease in winter precipitation at Shimla district for the respective period. P11-5 at Shimla district increased substantially in the 1980s but fell in the 1990s and 2000s with no recovery. Summer precipitation at districts Shimla and Dehradun also declined from the 1970s but passage of monsoon weakens northward and has little influence on runoff. Downstream, in the lower Sutlej, homogenous discharge records at Suni are not affected by artificial canal networks unlike Bhakra dam (see section 5.1) and reflect natural flows which are influenced by its headwaters upstream and by contemporary levels of precipitation. Volume of riverflow was enhanced on average by 85% from Khab to Suni as a result of additional flow from tributary rivers and direct precipitation predominantly from the summer monsoon. Despite increasing intensity and total amount of precipitation in the middle and lower reaches of the Sutlej, specific runoff at Suni was 0.261m and reflected a generally arid basin. Discharge at Suni decreased from maxima in the 1970s to a minimum in 2000, mimicking the fall of riverflow upstream at Khab, but declined by only 12% for the same period, 1972/81-1993/02 (figure 7-5). Discharge at Suni reflected both summer and winter precipitation. P6-9 at districts Shimla and Dehradun decreased to the 1980s before recovering slightly in the 1990s, P6-9 at Dehradun district increasing greatly in the 2000s.

Discharge in the lower Sutlej is influenced by waters upstream in the Tibetan portion of the basin and by meteorological conditions in the middle and lower reaches. Discharge in the lower Sutlej was examined without the influence of its headwaters in Tibet, which declined from the 1970s;

$$Q_{ls} = Q_s - Q_k$$

Discharge in the lower Sutlej ($Q_{ls}$) was calculated by subtracting annual totals of discharge in the lower Sutlej at Suni ($Q_s$) with annual totals of discharge in the upper Sutlej at Khab ($Q_k$). $Q_{ls}$ declined from the 1970s, reaching a minimum in the early 1980s, before rising from the 1990s to a maxima in 2002 (figure 7-5). Discharge increased by around 13% from 1972/81 to 1993/02, and by 26% from 1980/89 to 1993/02. Year-to-year fluctuations of $Q_{ls}$ mirrored the pattern of monsoon precipitation at Shimla district (correlation 0.45) which increased by around 6% from 1972/81-1993/02. Correlations were weaker between $Q_{ls}$ and summer monsoon at Dehradun district (correlation of 0.22). Winter precipitation at districts Shimla and Dehradun had a negative effect on $Q_{ls}$ (correlations of -0.25 and -0.16), wetter winters resulting in lower totals of annual runoff. Winter precipitation and summer precipitation had a negative relationship; drier winters generally strengthened the monsoon in the following summer and thus enhanced annual totals of discharge. Correlation between P11-5 and P6-9 at Shimla district was -0.31 and -0.18 between P11-5 and P6-9 at Dehradun district. An inverse relationship between winter precipitation and summer monsoon was also apparent in the central Himalaya by Dey and Kumar (1983).

## 7.2. Temporal variations of air temperature, precipitation and riverflow downstream at large dams

Riverflow at Tarbela is fed by its glacierised and less glacierised tributary basins upstream and thus has moderate interactions between temperature and precipitation variations. Q1-12 at Tarbela declined from its maxima in 1973 to a minimum in the 1980s, varying considerably from year-to-year in the 1990s and 2000s, declining by around 4% from the first to last decade of record, 1962/71–2000/09 (figure 7-6). Warm years generally enhanced runoff at Tarbela as its two largest and most glacierised tributaries, Hunza and Shyok, provided large quantities of runoff from melting of snow and ice. Warm summers 1973 and 1978 at Gilgit and Srinagar stand out as peak runoff years at Tarbela, with both years being peak runoff years at Hunza and Shyok. 1990 and 1994 were also particularly warm summers which enhanced riverflow. 1988 was ranked second maximum riverflow at Tarbela despite no peak flows from either of its tributaries. A large monsoon in 1988 penetrating from the east at Srinagar augmented riverflow in the lower reaches of the Indus where the volume of flow from Partab bridge (Indus below Astore) to Tarbela increased by around 52% (average increase around 41%).

To the east of the Indus basin, riverflow at Mangla in the near ice free Jhelum basin was influenced largely by winter precipitation derived from westerly disturbances. Q1-12 at Mangla mimicked the temporal variation of P11-5 at Srinagar, riverflow declining from the 1950 maxima to the 1970s before fluctuating along a rising trend to the 1990s, before falling considerably to 2000, overall flow increasing by around 15% between 1960-69 and 1991-2000 (figure 7-6). Summer precipitation at Srinagar also augmented riverflow, P6-9 at Srinagar increasing from the 1960s to maxima in the late 1980s and 1990s, before declining in 2000 with declining P11-5 and Q1-12. Summer air temperature at Srinagar had an inverse relationship with riverflow. T5-9 at Srinagar declined from the 1970s to 1990s as wetter winters and summers prevailed, increasing annual yields of runoff. A return to warmer summers in the late 1990s and 2000 led to reduced riverflow as a result of less precipitation. The third warmest summer of the series at Srinagar in 1971 was accompanied by a secondary minimum of P11-5 at Srinagar and a secondary minimum of Q1-12 at Mangla. Q1-12 in the Chenab basin at Marala followed similar year-to-year variations of discharge at Mangla. Annual total discharge at Marala which was largely driven by winter precipitation declined from the 1950s maxima to the 1970s, inclining to the 1990s, before falling in the 2000s by around 15% from the first to the last decade of discharge record; 1948/57-1997/06. Slightly more glacierised than its neighbouring Jhelum basin, discharge at Marala was not negatively influenced by summer air temperature variations. Warm summers in 1973 and 1978 at Srinagar appeared to enhance ice melt and riverflow when seasonal precipitation was subdued, but there was no overall influence between summer temperatures variations at Srinagar and annual discharge at Marala. Summer monsoon precipitation penetrating from the east influenced the runoff regime from June to September. From the late 1990s runoff at Marala declined considerably to a secondary minimum in 2001 (lowest since 1974) as winter and summer precipitation both declined.

## 7.3. Year-to-year variability of runoff

Year-to-year variability of annual runoff for a 21 year period (1980-2000) for which data is available for all gauging stations was highest in near ice free basins (table 6-1). Variability of annual runoff was highest at Khab (CV=0.24) where runoff was influenced by small amounts of winter precipitation and specific runoff was the lowest of any other basin (0.166m), which reflects arid conditions within the basin. Mangla and Astore shared the same variability of runoff (CV=0.22) where runoff was also influenced mainly by winter precipitation. Variability of annual runoff in an ice free/ near ice free basin reflects solely variability of precipitation as there is little ice melt to offset reduced runoff in warmer years when precipitation is low (e.g. Tollan 1972, Tvede 1982, Collins 1990). Annual variability of runoff at Khab over a 21 year period (CV=0.24) was higher but similar to variability of P1-12 at Shimla district (CV=0.17) over the same period, 1980-2000. Variability of discharge at Astore and Mangla (CV=0.22) had similar variations to annual precipitation at Srinagar district (CV=0.23) over a 21 year period, whilst precipitation variability at Astore was CV=0.27. Runoff variability in the near ice free Sutlej basin decreased downstream to a minimum at Bhakra where CV was the lowest of any basin over a 21 year period 1980-2000 (CV=0.11). Also influenced by precipitation, runoff variability at Bhakra is reduced as it receives runoff influenced by winter precipitation from the north of its catchment in spring and early summer and runoff from the summer monsoon in the middle and lower reaches of it basin from mid to late summer. Runoff is also artificially enhanced at Bhakra from the Beas-Sutlej link canal at Dehar which may compensate runoff in drier years by passage of more water through its canal system.

Basins with varying proportions of ice covered area in the UIB had a wide range of CV between 1980 and 2000 but were lower than CV in near ice free basins. Annual variability of runoff in the large ice covered Shyok basin was influenced predominantly by air temperature fluctuations and had variability which was marginally lower than in near ice free basins (CV=0.19). Variability of energy fluctuations tend to be lower than precipitation variability (e.g. Collins 1990), thus ice free basins have higher variability than ice dominated basins. A large proportion of the Shyok catchment lies in arid Ladakh and Tibet where precipitation totals are low and are highly variable from year-to-year. The runoff component from precipitation is small (Mukhopadhyay and Khan 2014b) and therefore does not provide enough streamflow to moderate runoff during cooler years when icemelt runoff is subdued. In the western Karakoram valley floors are also arid, receiving typically less than 200mm annually, but precipitation increases with altitude and is less variable than in the eastern Karakoram (Immerzeel et al 2009). Runoff at Hunza reflects air temperature fluctuations, warmer summers provide higher runoff totals over wet winters, but winter precipitation at higher altitudes provides some moderation of runoff during cooler years and thus reduces runoff variability.

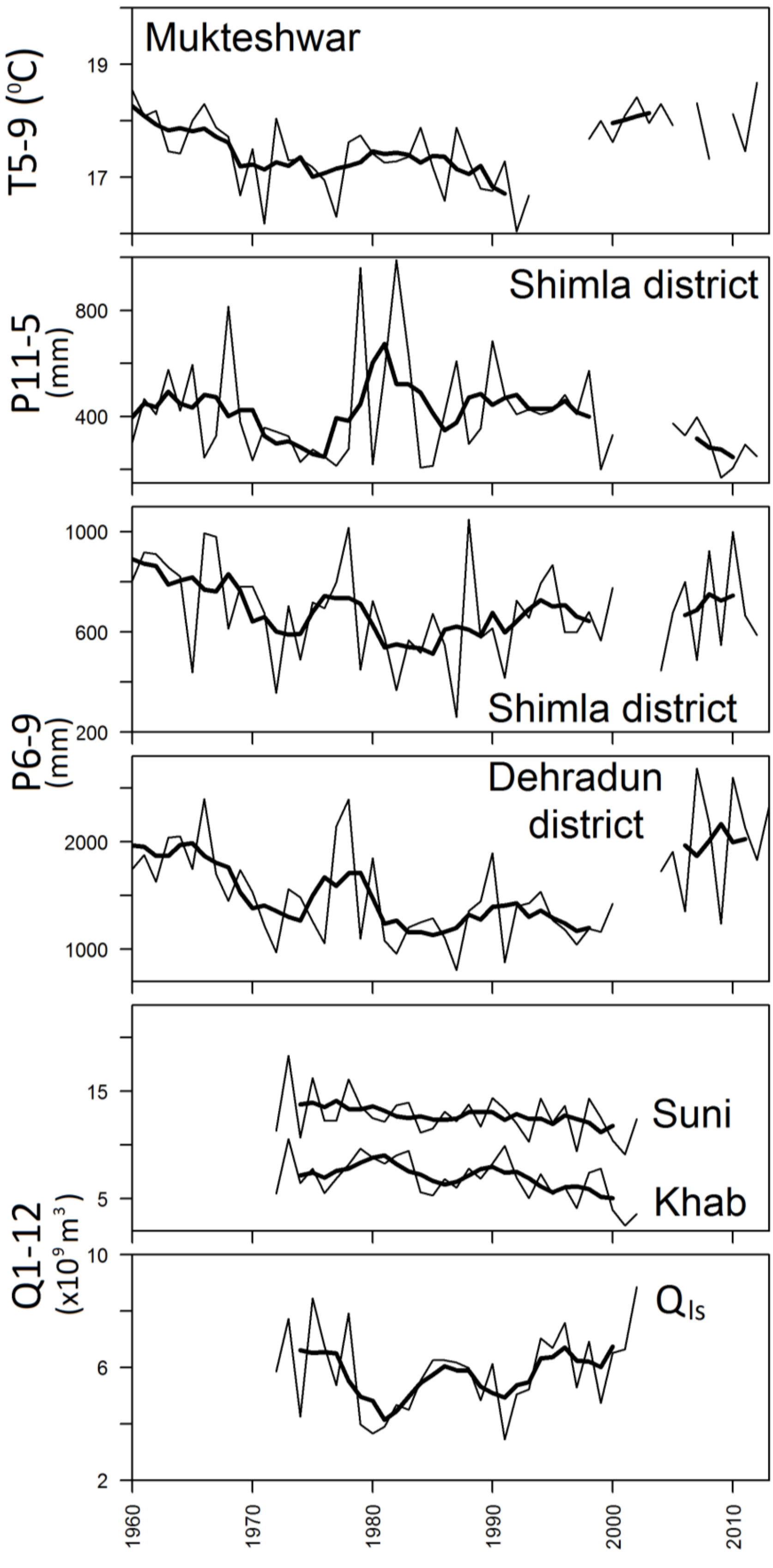


**Figure 7-5: Year-to-year variations of mean summer air temperature at Mukteshwar, totals of winter precipitation at Shimla district, totals of summer precipitation at districts Shimla and Dehradun, annual totals of discharge at Suni and Khab, and calculated discharge from the lower Sutlej ($Q_{ls}$) from 1960-2013.**

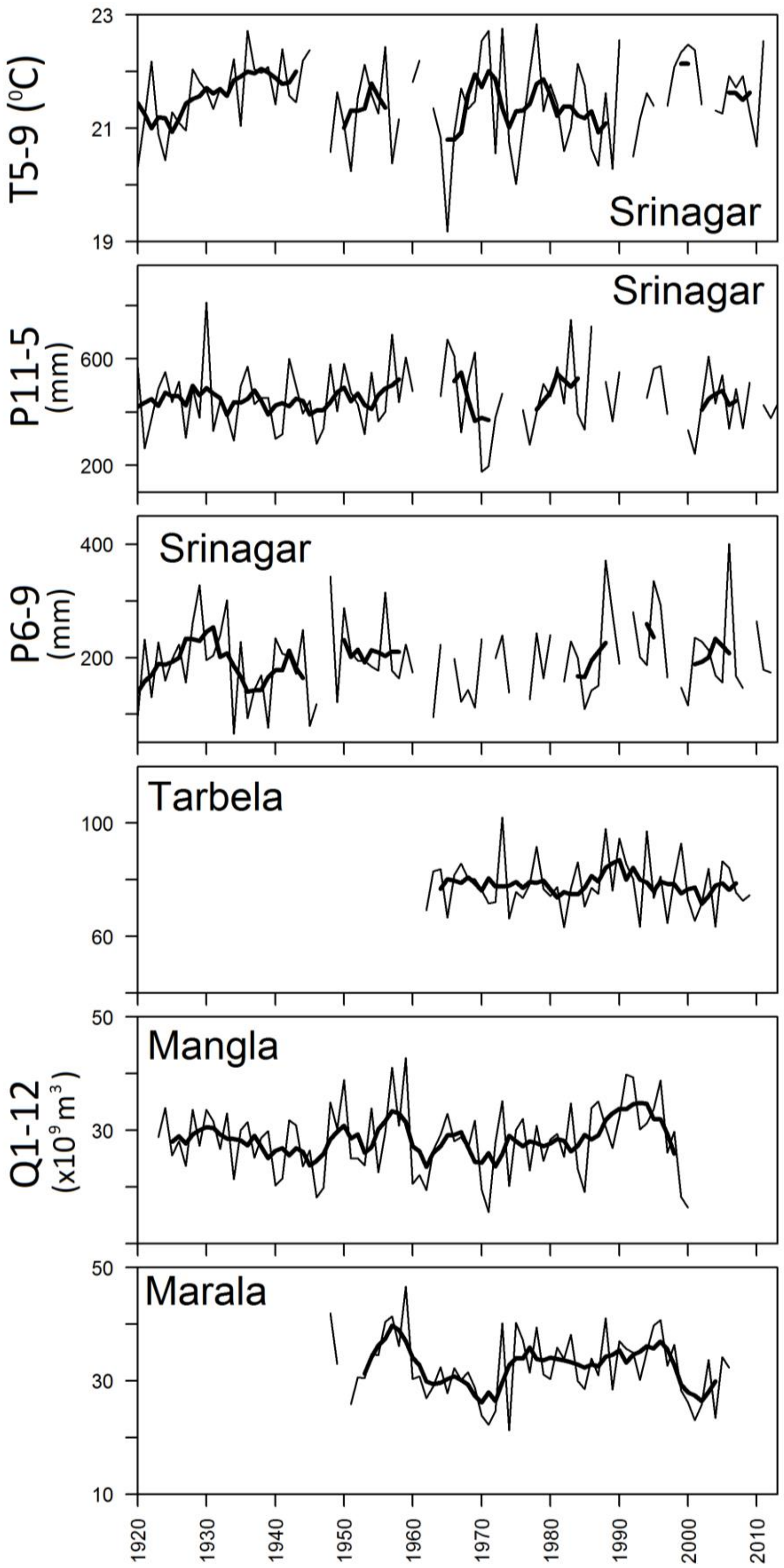


**Figure 7-6: Year-to-year variations of mean summer air temperature at Srinagar, totals of winter precipitation and summer precipitation at Srinagar, and annual totals of discharge at Tarbela (Indus), Mangla (Jhelum), and Marala (Chenab) from 1920-2013.**

In the intermediate ice covered Gilgit basin runoff was mainly influenced by winter precipitation. CV of Q1-12 at Gilgit (CV=0.12) was substantially less than P1-12 at Gilgit (CV=0.36). Warmer years enhanced icemelt runoff, compensating for reduced runoff over the ice free areas which received less precipitation. For a similar period of record (1981-2005) CV of annual runoff in glacierised Alpine basins was considerably less; ranging from CV=0.099 in an intermediate glacierised Alpine basin to CV=0.120 – CV=0.142 in highly glacierised basins (Collins 2007). Basins in the Karakoram and Tibet have low annual totals of precipitation at elevations between 1000-3000m a.s.l, and precipitation increases above ~3500m – 4000m a.s.l (Immerzeel et al 2012). The lower ice free portion of a Karakoram-Tibetan basin is essentially hydrologically inactive and does not provide enough runoff to moderate riverflow in cooler years when icemelt is reduced. Alpine basins in Switzerland and USA efficiently moderate runoff as precipitation is received at all elevations (e.g. Collins 1990, Chen and Ohmura 1990, Fountain and Tangborn 1985).

Bhakra had the lowest year-to-year variability of the large dams between 1980-2000, followed by Tarbela, Marala and Mangla (Table 6-1). Over a 39 year period (1962-2000) Bhakra and Tarbela shared the lowest variability (CV=0.12), followed again by Marala and Mangla, but over 43 years (1962-2004) Tarbela had the lowest Variability (CV=0.12) compared with CV=0.14 at Bhakra. Discharge at Tarbela is moderated by receiving three distinct runoff components which are non-synchronous in yearly variation;

1) Icemelt runoff from its glacial tributaries (Hunza and Shyok)
2) Snowmelt runoff from its less glacierised tributaries (Gilgit and Astore)
3) Monsoon runoff from tributaries below Astore

In warm summers following drier winters runoff from less glacierised snowmelt catchments (Gilgit, Astore) is reduced, but riverflow in glacierised basins (Shyok, Hunza) is enhanced. Likewise, wet years followed by cool summers produce less runoff from glacierised basins as energy availability to melt ice is reduced, but runoff from less glacierised basins which have a larger ice-free area will compensate for the icemelt deficit through increased precipitation runoff. Covering an area of 172,159km$^2$ meteorological conditions within the UIB vary spatially. Warm summers from part of the Indus may coincide with increased precipitation from a different region of the Indus which may lead to enhanced runoff downstream. Combined, tributaries Hunza, Shyok, Gilgit and Astore contribute around 44% of total discharge at Tarbela dam. In the warm

1970s when discharge was at its maxima in the Hunza basin, large amounts of icemelt contributed on average around 15.5% of the flow at Tarbela, decreasing to around 12% in the 1990s and 2000s as declining summer air temperatures in the Karakoram reduced icemelt. Its neighbouring Gilgit basin stayed consistent, as increasing winter precipitation compensated for reduced icemelt, contributing around 11.6% in the 1960s and 1990s, increasing by around 1% in the 2000s. At Astore the percentage contribution of riverflow to Tarbela was small but increased from around 5% to a maximum of 7.7% in 1996 as precipitation increased runoff. Despite the percentage of runoff contribution reducing to a minimum in the late 1980s at Hunza, Shyok and Gilgit, riverflow at Tarbela increased. Riverflow at Shyok increased from the 1990s as increased temperatures along the Tibetan Plateau enhanced icemelt, compensating for reduced riverflow at Hunza and stabilising riverflow at Tarbela. Khattak et al (2011) reported a non-significant decline of June-August riverflow at Besham and suggested water availability at Tarbela will be impacted. This study suggests that water resources in the UIB at Tarbela are stable and on an annual basis are moderated by seasonal variations. Low variability of annual runoff at Tarbela over a period of 43 years suggests stable riverflow for the foreseeable future, and should glacier melt runoff continue to decline at Hunza runoff from precipitation fed basins in the middle and lower reaches will compensate for this deficit.

## 7.4. Relationship between climatic variables

### 7.4.1. Correlation coefficients of summer precipitation time series

Summer precipitation from June to September was reasonably correlated between stations in the UIB (table 7-1) despite differing absolute quantities of precipitation (table 5-1). All stations in the UIB followed a similar pattern of summer precipitation; decreasing in the 1960s, increasing to the 1990s, and falling slightly in the early 2000s. Highest correlation coefficients were between Gilgit and Bunji ($r = 0.72$) despite a difference in elevation of around 800m, followed closely by $r = 0.70$ between Chilas and Astore, both stations being in close proximity but having an elevation difference of over 900m. Weakest correlations were obtained between Chilas and Skardu ($r = 0.48$) where the distance between stations was around 140km. An average correlation was shared between Skardu and Gilgit over an 85 year overlap between 1900 and 1999 ($r = 0.51$) but the correlation strengthened to $r = 0.65$ from 1947 to 1999, accounting for the site changes which occurred during the partition of India in 1947. Correlations between P6-

9 reduced with greater distances between climatic stations. Correlations between P6-9 in the Indus were weakly or negatively related to those in the western Himalayas, suggesting summer precipitation in the UIB is not entirely monsoon, as a strong monsoon in the western Himalayas appears to reduce summer precipitation in the Karakoram. Correlations were negative between some climatic stations in the western Himalaya and Gilgit, strongest correlations being -0.11 with Chakrata and -0.10 with Shimla. P6-9 at Chilas had strong negative correlations with Chakrata (r= -0.43) and Shimla (r= -0.33) but had short overlapping records, 11 and 30 years respectively. Rampur had a positive correlation with P6-9 at Gilgit (r= 0.30) and a strong negative relationship with Chilas (r= -0.88) but only had an overlap of 6 years with each station which is not representative of their true relationship. Summer precipitation at Srinagar had reasonable correlations with climate stations in the UIB and had slightly higher correlations than the Srinagar district time series. Correlations between Srinagar and western Himalayan stations were weak with no relationship being established with Shimla (r= -0.03), Shimla district (r= 0.08), and Dehradun strict (r= 0.01).

Correlations of summer precipitation between climate stations in the western Himalaya were generally stronger but were less consistent. Long time series at Shimla correlated highly with its district series (r= 0.82) as did the long series at Dehradun with both Shimla (r= 0.62) and Shimla district (r= 0.66). Low altitude climate station Ludhiana had fairly weak correlations with its neighbouring stations, but had a high correlation with Bhakra (r= 0.78) with just 6 year overlap of precipitation record. P6-9 at Bhakra had a strong positive correlation with Shimla (r= 0.89) but only had a 5 year overlap, whilst a 27 year overlap with Shimla district and Dehradun district had mixed correlations, r= 0.79 and 0.24, respectively. Low lying meteorological station Chakrata had reasonable correlations Shimla and Dehradun.

### 7.4.2. Correlation coefficients of winter precipitation time series

Correlations of winter precipitation from November to May in the UIB were generally stronger than those correlations with summer precipitation (table 7-2). Strongest correlations were again between Gilgit and Bunji (r= 0.74), P11-5 increasing from the 1980s-2000s in the Indus basin. Weakest correlation was between Gilgit and Skardu over a 76 year period between 1901 and 1999 (r= 0.40), increasing to r= 0.45 for a shorter period 1947-1999 after site location changes. Both Srinagar and Srinagar district had weaker correlations with P11-5 in the Indus than with P6-9. Astore had the highest correlation of any station in the UIB with Srinagar (r= 0.46) being located within a short distance and on the border of the western Himalaya. Srinagar had a weak correlation with Srinagar district (r= 0.39), both following contrasting patterns of precipitation in the first half of the 20$^{th}$ century. Correlations between P11-5 from meteorological stations in the Indus and western Himalaya were weakly related, correlations turning negative at some stations. Skardu had a negative correlation with Bhakra (r= -0.36) over a short 7 year overlap between records and a negative correlation with Chakrata (r= -0.17) and Dehradun district (r= -0.25) over 30 and 75 years. Bunji, Gilgit and Astore were also negatively related to Bhakra and Chakrata, Astore only having a 9 year overlap with Chakrata.

Correlations of P11-5 in the western Himalaya were heterogeneous. Correlations between Shimla, Shimla district, Dehradun district and Chakrata were fairly strong, all stations being located in mountain regions. Correlations of P11-5 were weakest with Bhakra on the Punjab plains, r= -0.07 with Shimla, r = 0.18 with Shimla district, r= 0.34 with Dehradun district. Ludhiana, also on the Punjab plains, had surprisingly good correlations with neighbouring mountain stations, the lowest being r= 0.57 with Shimla. Srinagar and Srinagar district had fairly weak/average correlations between both western Himalayan and UIB climate stations but had overall stronger correlations with the western Himalaya.

### 7.4.3. Correlation coefficients of summer air temperature time series

Summer air temperature in the UIB which declined from around the 1960s was less spatially variable than P6-9 and P11-5 as suggested by higher correlations (table 7-3). Highest correlation existed between Gilgit and Astore, and Gilgit and Chilas (r= 0.88), all stations being in close proximity. Weakest correlation in the Indus basin were between Bunji and Astore (r= 0.72) but were still strongly correlated. T5-9 from Tibetan climate stations Hotan and Kashi followed a similar pattern to temperatures in the UIB; declining to the 1990s but diverging to warm 2000s, having reasonable correlations with climate stations in the Indus despite large distances between stations. Weakest correlations were between Tibetan stations and Bunji. Correlations between Indus climate stations and Tibetan station Shiquanhe weakened being situated further east but had high correlations with other Tibetan stations Hotan and Kashi. T5-9 at Srinagar correlated well with summer temperatures from stations in the UIB, correlations being slightly weaker than July-September air temperature correlations up to the year 2000 by Fowler and Archer (2006). Highest correlation was between Srinagar and Astore (r= 0.78) both stations being at a similar elevation and within a close distance. Srinagar had weaker correlations with stations in the western Himalayas, having an average correlation with neighbouring Jhelum (r= 0.41) and Amritsar (r= 0.39), poor correlation with Sialkot (r= 0.13) and Ludhiana (r= 0.17), and turning negative with Mukteshwar (r= -0.09) in the western-central Himalaya.

Correlations of T5-9 between Himalayan climate stations varied, stations being located at mixed elevations in mountainous and non-mountainous regions. Climate stations at low altitudes south of the main Himalaya were closely related; Jhelum, Sialkot, Amritsar and Ludhiana all having good correlations, temperatures declining from around the 1960s/70s and warming from the 1990s. Shimla, located in the main Himalaya, had high correlations with lowland stations Jhelum and Sialkot but only average correlations with Amritsar and Ludhiana. Correlations between low altitude stations were reduced with Mukteshwar, lowest correlation being between Mukteshwar and Amritsar (r= 0.34), but a high correlation existed with mountain stations Shimla and Mukteshwar (r= 0.71). Weak negative correlations existed between T5-9 from stations in the UIB and the Himalaya. A 46 year overlap between UIB stations and Mukteshwar had correlations in the range of r= -0.11 to -0.26. Weak positive correlations existed between Tibetan climate stations and Mukteshwar, but were negative with shorter overlapping records with Ludhiana and Shimla. T5-9 at Amritsar had reasonable correlations with climate stations in the UIB over 46-48 years of overlapping records, summer temperatures at Amritsar cooling from the 1950s.

**Table 7-1: Correlation coefficients of summer precipitation between meteorological stations used in this study. Parentheses indicate number of overlapping years between correlations.**

| | Gilgit | Skardu | Bunji | Astore | Chilas | Srinagar | Srinagar district | Ludhiana | Bhakra | Shimla | Shimla district | Dehradun district |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Skardu** | 0.51 (85) | | | | | | | | | | | |
| **Bunji** | 0.72 (61) | 0.49 (46) | | | | | | | | | | |
| **Astore** | 0.61 (60) | 0.68 (45) | 0.61 (59) | | | | | | | | | |
| **Chilas** | 0.58 (60) | 0.48 (46) | 0.54 (60) | 0.70 (59) | | | | | | | | |
| **Srinagar** | 0.43 (105) | 0.47 (75) | 0.43 (50) | 0.42 (49) | 0.49 (49) | | | | | | | |
| **Srinagar district** | 0.34 (101) | 0.48 (83) | 0.29 (53) | 0.40 (52) | 0.38 (52) | 0.66 (93) | | | | | | |
| **Ludhiana** | 0.11 (71) | 0.37 (71) | 0.18 (32) | 0.21 (31) | 0.31 (31) | 0.34 (85) | 0.15 (84) | | | | | |
| **Bhakra** | 0.22 (30) | 0.54 (16) | 0.16 (30) | 0.22 (30) | 0.32 (30) | 0.30 (27) | 0.31 (23) | 0.78 (6) | | | | |
| **Shimla** | -0.10 (88) | 0.11 (70) | 0.01 (31) | 0.29 (30) | -0.33 (30) | -0.03 (94) | 0.02 (83) | 0.14 (112) | 0.89 (5) | | | |
| **Shimla district** | -0.07 (105) | 0.14 (84) | 0.17 (57) | 0.19 (56) | 0.01 (56) | 0.08 (97) | 0.15 (105) | 0.28 (84) | 0.79 (27) | 0.82 (83) | | |
| **Dehradun district** | -0.03 (106) | 0.19 (84) | 0.07 (58) | 0.14 (57) | -0.10 (57) | 0.01 (97) | 0.05 (105) | 0.32 (84) | 0.24 (27) | 0.62 (83) | 0.66 (108) | |
| **Chakrata** | -0.11 (60) | 0.03 (50) | 0.13 (13) | -0.22 (11) | -0.43 (11) | 0.20 (60) | 0.14 (64) | 0.30 (64) | | 0.47 (64) | 0.54 (64) | 0.61 (64) |
| **Rampur** | 0.30 (6) | | 0.69 (6) | -0.05 (6) | -0.88 (6) | -0.42 (3) | -0.86 (3) | | 0.05 (6) | | 0.56 (3) | 0.54 (3) |

**Table 7-2: Correlation coefficients of winter precipitation between meteorological stations used in this study. Parentheses indicate number of overlapping years between correlations.**

| | Gilgit | Skardu | Bunji | Astore | Chilas | Srinagar | Srinagar district | Ludhiana | Bhakra | Shimla | Shimla district | Dehradun district |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Skardu** | 0.40 (76) | | | | | | | | | | | |
| **Bunji** | 0.74 (59) | 0.50 (42) | | | | | | | | | | |
| **Astore** | 0.66 (59) | 0.70 (43) | 0.66 (58) | | | | | | | | | |
| **Chilas** | 0.73 (60) | 0.45 (43) | 0.70 (60) | 0.71 (59) | | | | | | | | |
| **Srinagar** | 0.27 (103) | 0.24 (65) | 0.34 (48) | 0.46 (47) | 0.45 (49) | | | | | | | |
| **Srinagar district** | 0.27 (98) | 0.01 (74) | 0.25 (51) | 0.34 (50) | 0.27 (52) | 0.39 (93) | | | | | | |
| **Ludhiana** | 0.11 (98) | -0.02 (58) | 0.54 (28) | 0.23 (27) | 0.30 (29) | 0.36 (84) | 0.33 (80) | | | | | |
| **Bhakra** | -0.17 (29) | -0.36 (16) | -0.21 (29) | -0.11 (29) | -0.02 (29) | 0.11 (22) | 0.09 (22) | | | | | |
| **Shimla** | 0.02 (96) | -0.09 (62) | 0.06 (31) | -0.07 (30) | 0.17 (32) | 0.33 (86) | 0.25 (83) | 0.69 (110) | -0.07 (6) | | | |
| **Shimla district** | 0.10 (102) | -0.01 (76) | -0.03 (55) | 0.03 (54) | 0.24 (56) | 0.41 (95) | 0.11 (103) | 0.57 (80) | 0.18 (18) | 0.69 (83) | | |
| **Dehradun district** | -0.01 (103) | -0.25 (76) | -0.07 (56) | -0.19 (55) | 0.01 (57) | 0.42 (96) | 0.34 (103) | 0.60 (80) | 0.34 (25) | 0.73 (83) | 0.63 (107) | |
| **Chakrata** | -0.18 (45) | -0.17 (30) | -0.13 (11) | -0.36 (9) | -0.09 (11) | 0.39 (46) | 0.37 (49) | 0.65 (49) | | 0.75 (49) | 0.52 (49) | 0.88 (49) |
| **Rampur** | 0.32 (5) | | 0.32 (5) | 0.50 (5) | 0.24 (5) | 0.25 (5) | | | 0.29 (5) | | | |

**Table 7-3: Correlation Coefficients of summer air temperature time series from meteorological stations used in this study. Parentheses indicate number of overlapping years between correlations.**

| | Gilgit | Bunji | Astore | Chilas | Hotan | Kashi | Shiquanhe | Srinagar | Jhelum | Sialkot | Amritsar | Ludhiana | Shimla |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Bunji** | 0.78 (53) | | | | | | | | | | | | |
| **Astore** | 0.88 (52) | 0.72 (52) | | | | | | | | | | | |
| **Chilas** | 0.88 (51) | 0.75 (51) | 0.84 (50) | | | | | | | | | | |
| **Hotan** | 0.58 (51) | 0.24 (51) | 0.66 (50) | 0.52 (49) | | | | | | | | | |
| **Kashi** | 0.47 (51) | 0.26 (51) | 0.58 (50) | 0.32 (49) | 0.90 (54) | | | | | | | | |
| **Shiquanhe** | 0.32 (33) | 0.05 (33) | 0.36 (32) | 0.14 (31) | 0.70 (33) | 0.71 (32) | | | | | | | |
| **Srinagar** | 0.68 (47) | 0.41 (47) | 0.78 (46) | 0.69 (45) | 0.67 (53) | 0.59 (48) | 0.55 (30) | | | | | | |
| **Jhelum** | 0.01 (51) | -0.06 (51) | 0.10 (50) | 0.03 (49) | 0.29 (59) | 0.34 (51) | 0.49 (31) | 0.41 (45) | | | | | |
| **Sialkot** | -0.50 (19) | -0.43 (19) | -0.34 (18) | -0.57 (17) | -0.41 (64) | 0.05 (18) | 0.32 (16) | 0.13 (29) | 0.73 (23) | | | | |
| **Amritsar** | 0.27 (48) | 0.40 (48) | 0.41 (47) | 0.36 (46) | 0.15 (56) | 0.21 (50) | 0.15 (32) | 0.39 (55) | 0.59 (56) | 0.63 (14) | | | |
| **Ludhiana** | -0.28 (15) | 0.04 (15) | -0.28 (14) | -0.26 (14) | -0.04 (28) | -0.15 (17) | -0.08 (14) | 0.17 (77) | 0.80 (23) | 0.75 (18) | 0.84 (26) | | |
| **Shimla** | -0.35 (25) | -0.42 (25) | -0.38 (24) | -0.47 (23) | 0.05 (40) | -0.11 (28) | 0.22 (20) | 0.07 (88) | 0.69 (32) | 0.77 (22) | 0.46 (36) | 0.48 (91) | |
| **Mukteshwar** | -0.23 (46) | -0.13 (46) | -0.11 (45) | -0.26 (44) | 0.08 (59) | 0.24 (48) | 0.18 (30) | -0.09 (103) | 0.55 (56) | 0.46 (23) | 0.34 (56) | 0.41 (77) | 0.71 (88) |

### 7.4.4. Correlation coefficients between precipitation and air temperature

Correlation coefficients between precipitation variables and summer air temperature for respective stations are provided in tables 7-4 and 7-5. Wet winters were generally accompanied by cool summers the following year in the UIB, correlations ranging from r= -0.30 to -0.43 (table 7-4). Relatively dry winters in 1960s/1970s mirrored warm summers, temperatures cooling to the 1990s as precipitation increased to its maxima. P11-5 and T5-9 in the western Himalaya were, however, unrelated, except for at Srinagar, correlations between P11-5 and T5-9 at Ludhiana and Shimla being r= -0.08, respectively. Summer precipitation was also negatively related to summer air temperature at all stations except Shimla, where no relationship existed between P6-9 and T5-9 (table 7-5). Highest correlation between P6-9 and T5-9 was at Ludhiana (r= -0.55), warmer years in the early 1900 and 1930s reflected reduced precipitation, summer temperatures reducing to the 1960s with increasing precipitation. Only a weak negative relationship between P6-9 and T5-9 was established at Srinagar over a period of 103 years (r= -0.19), warmer summers in the 1930s/40s and 1970s inversely matched drier summers, temperatures reduced slightly in the wetter 1990s but recovered into the 2000s following consecutive dry summers.

**Table 7-4: Correlation coefficients between P11-5 and T5-9 from respective stations. Parentheses indicate number of overlapping years between correlations.**

| | Gilgit T5-9 | Bunji T5-9 | Astore T5-9 | Chilas T5-9 | Srinagar T5-9 | Ludhiana T5-9 | Shimla T5-9 |
|---|---|---|---|---|---|---|---|
| Gilgit P11-5 | -0.30 (53) | | | | | | |
| Bunji P11-5 | | -0.41 (52) | | | | | |
| Astore P11-5 | | | -0.43 (52) | | | | |
| Chilas P11-5 | | | | -0.38 (51) | | | |
| Srinagar P11-5 | | | | | -0.38 (101) | | |
| Ludhiana P11-5 | | | | | | -0.08 (95) | |
| Shimla P11-5 | | | | | | | -0.08 (102) |

**Table 7-5: Correlation coefficients between P6-9 and T5-9 from respective stations. Parentheses indicate number of overlapping years between correlations.**

| | Gilgit T5-9 | Bunji T5-9 | Astore T5-9 | Chilas T5-9 | Srinagar T5-9 | Ludhiana T5-9 | Shimla T5-9 |
|---|---|---|---|---|---|---|---|
| **Gilgit P6-9** | -0.37 (53) | | | | | | |
| **Bunji P6-9** | | -0.41 (53) | | | | | |
| **Astore P6-9** | | | -0.28 (52) | | | | |
| **Chilas P6-9** | | | | -0.42 (51) | | | |
| **Srinagar P6-9** | | | | | -0.19 (103) | | |
| **Ludhiana P6-9** | | | | | | -0.55 (94) | |
| **Shimla P6-9** | | | | | | | 0.04 (103) |

### 7.4.5. Relationship between climatic variables and total annual riverflow

The extent to which climatic variation influences runoff from glacierised basins can be assessed through examination of correlation coefficients from nearby meteorological stations. The markedly similar variations of T5-9 in the UIB with Q1-12 from glacierised basins Shyok and Hunza suggests energy availability exerts a strong influence on total annual runoff (table 7-6 figure 7-7). Form and timing of precipitation also influences runoff, but has a complex relationship. Wet winters add a component of runoff in highly glacierised basins Shyok and Hunza but the ice free area over which provides runoff is limited and arid. Particular wet winters retard the upward rise of the transient snowline in early spring and summer and thus limit the exposure length of clean ice to melting. P11-5 at Gilgit had a weak but positive relationship with Q1-12 at Shyok and Hunza, r= 0.22 and 0.17, respectively, but correlations of Q1-12 at Shyok and Hunza with P11-5 at other neighbouring stations was reduced, turning negative with runoff at Hunza (table 7-7). Summer precipitation had an inverse relationship with runoff at Shyok and Hunza although the relationship was stronger with Hunza (table 7-8) (figure 7-7). Wet summers reduce incoming solar radiation with the associated increase of cloud cover, but precipitation in the form of snow increases the albedo of the glacier surface causing recessive flows (e.g. Oerlemans and Klok 2004). Increasing summer precipitation at Gilgit since around the 1980s reduced runoff at Hunza. Shyok being less influenced by summer precipitation followed patterns of T5-9 in Tibet, where correlations are strong. As basin ice cover reduces, the relationship between runoff and climatic variability is reversed, precipitation augments riverflow and energy availability has little influence (figure 7-7). Q1-12 at Gilgit and Astore had a negative relationship with T5-9 in the Indus basin, the correlation being more negative at Astore where basin ice coverage is smaller. Winter precipitation provided explanation of riverflow trends at Gilgit and Astore, correlation with P11-5 from their catchment station being r= 0.56 and 0.67, respectively. Summer precipitation augmented riverflow in the least ice covered Astore basin, being mildly influenced by the Indian monsoon. Correlation between Astore Q1-12 and Astore P6-9 was r= 0.22, but increased to r= 0.51 with Srinagar P6-9, Srinagar receiving a larger proportion of monsoon precipitation. Total annual runoff in the intermediate ice covered Gilgit basin was reduced during wet summers, fresh snowfall raising the albedo of the glacier surface and retarding ablation of underlying ice, but was generally less negative than the highly glacierised Shyok and Hunza basin (table 7-8 and figure 7-7).

In the low glacierised Sutlej basin, runoff at Khab was predominantly controlled by winter precipitation, runoff being positively related to winter precipitation from its neighbouring meteorological stations (table 7-10), highest correlations being with P11-5 at Shimla (r= 0.50) and Shimla district (r= 0.38). Summer air temperature at Mukteshwar, which indicates energy availability for melting of ice, had a negative relationship with runoff at Khab over a 30 year period (r= -0.24) (table 7-9), warmer summers generally reducing totals of winter precipitation (table 7-4). Surprisingly summer precipitation at Shimla and Shimla district were negatively related to runoff at Khab (table 7-11), not because of increased surface albedo as seen in glacierised basins in the UIB, but because stronger monsoon seasons affected the passage of westerlies the following winter, as discussed in section 7.1.2 ( and seen by Dey and Kumar 1983). Weak winter westerlies in the western Himalayas generally enhanced the passage of the monsoon during summer, while a wet winter weakened the eastern monsoon the following summer. Correlation between Shimla district P11-5 and P6-9 was r= -0.36. The relationship between summer monsoon precipitation and total annual discharge strengthened with distance downstream on the Sutlej as the strength of the monsoon increased (table 7-11). For example; the correlation between annual discharge and P6-9 at Shimla district increased from r= -0.05 at Khab to r= 0.18 at Rampur and Luhri, to r= 0.26 and r= 0.55 at Suni and Bhakra. P6-9 at Shimla had a negative relationship with Q1-12 at Rampur, Luhri and Suni, but had a short overlap of records between 14 and 20 years. Likewise, correlations between P6-9 at Ludhiana and Q1-12 were weak with the common period of record being only 13-21 years. The relationship between winter precipitation and discharge decreased downstream as snow coverage and westerlies weakened (table 7-10). From overlapping periods between 29 and 38 years the correlation of discharge and P11-5 at Shimla district reduced steadily from r= 0.38 at Khab, r= 0.24 at Rampur, r= 0.21 and r= 0.17 at Luhri and Suni, to almost no correlation at Bhakra, r= 0.02 (table 7-10).

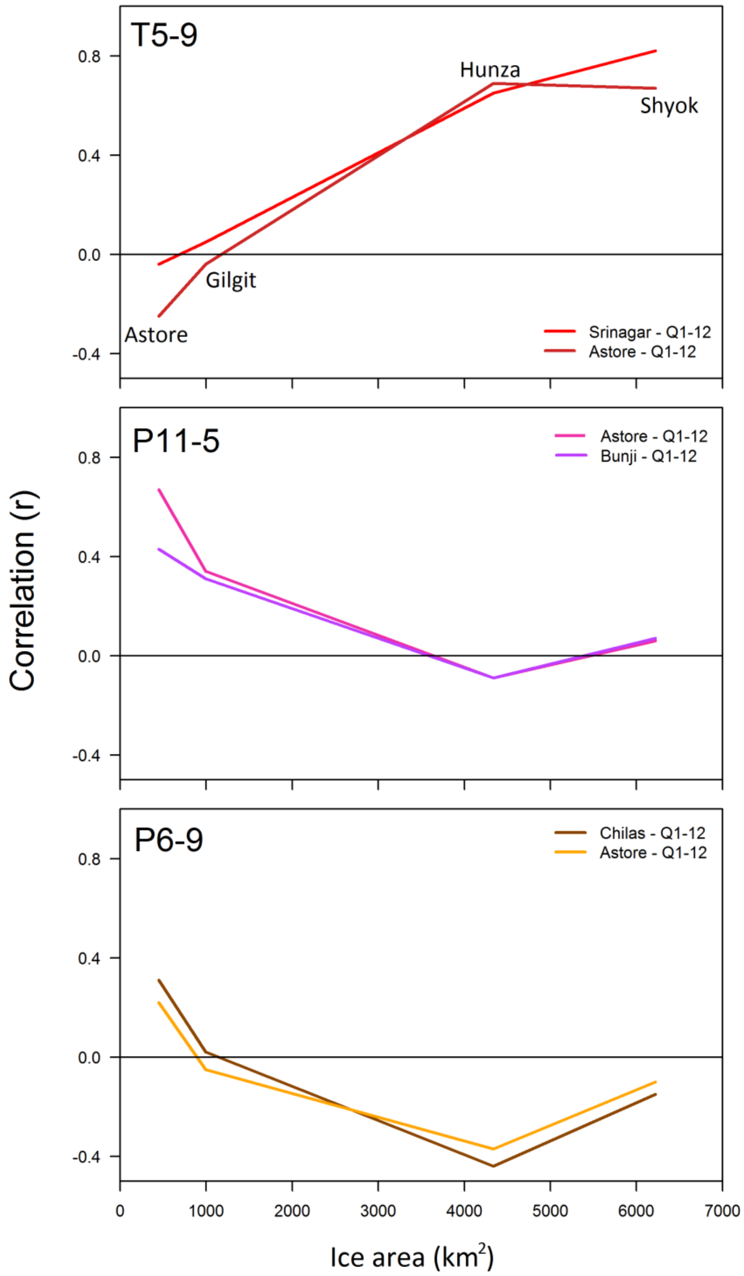


**Figure 7-7: Correlation coefficients between climatic variables and discharge from basins in the UIB as a function of total ice area of a basin**

**Table 7-6: Correlation coefficients between Q1-12 from basins in the UIB and T5-9. Parentheses indicate number of overlapping years between records.**

| | Gilgit T5-9 | Bunji T5-9 | Astore T5-9 | Chilas T5-9 | Hotan T5-9 | Kashi T5-9 | Shiquanhe T5-9 | Srinagar T5-9 |
|---|---|---|---|---|---|---|---|---|
| **Shyok Q1-12** | 0.56 (33) | 0.35 (33) | 0.67 (33) | 0.55 (31) | 0.72 (31) | 0.68 (32) | 0.62 (13) | 0.82 (30) |
| **Hunza Q1-12** | 0.68 (39) | 0.67 (39) | 0.69 (39) | 0.61 (37) | 0.31 (38) | 0.43 (38) | 0.27 (20) | 0.65 (36) |
| **Gilgit Q1-12** | -0.02 (38) | -0.11 (38) | -0.04 (38) | -0.14 (43) | 0.04 (44) | 0.13 (44) | -0.09 (25) | 0.05 (42) |
| **Astore Q1-12** | -0.20 (32) | -0.36 (32) | -0.25 (32) | -0.30 (30) | -0.23 (30) | -0.21 (31) | 0.84 (12) | -0.04 (29) |
| **Besham Q1-12** | 0.37 (29) | 0.12 (29) | 0.31 (29) | 0.24 (27) | 0.43 (29) | 0.45 (29) | 0.66 (17) | 0.44 (27) |

**Table 7-7: Correlation coefficients between Q1-12 from basins in the UIB and P11-5. Parentheses indicate number of overlapping years between records.**

| | Gilgit P11-5 | Skardu P11-5 | Bunji P11-5 | Astore P11-5 | Chilas P11-5 | Srinagar P11-5 |
|---|---|---|---|---|---|---|
| **Shyok Q1-12** | 0.22 (33) | 0.04 (26) | 0.07 (32) | 0.06 (33) | 0.11 (33) | -0.20 (25) |
| **Hunza Q1-12** | 0.17 (39) | -0.20 (33) | -0.09 (38) | -0.09 (39) | -0.10 (39) | -0.18 (31) |
| **Gilgit Q1-12** | 0.56 (39) | 0.31 (32) | 0.31 (39) | 0.34 (39) | 0.41 (39) | 0.31 (31) |
| **Astore Q1-12** | 0.45 (32) | 0.48 (25) | 0.43 (32) | 0.67 (32) | 0.63 (32) | 0.52 (24) |
| **Besham Q1-12** | 0.34 (29) | 0.31 (28) | 0.06 (28) | 0.36 (29) | 0.28 (29) | 0.24 (23) |

**Table 7-8: Correlation coefficients between Q1-12 from basins in the UIB and P6-9. Parentheses indicate number of overlapping years between records.**

| | Gilgit P6-9 | Skardu P6-9 | Bunji P6-9 | Astore P6-9 | Chilas P6-9 | Srinagar P6-9 |
|---|---|---|---|---|---|---|
| **Shyok Q1-12** | -0.07 (33) | -0.08 (26) | -0.18 (33) | -0.10 (33) | -0.15 (33) | -0.02 (28) |
| **Hunza Q1-12** | -0.34 (39) | -0.26 (33) | -0.35 (39) | -0.37 (39) | -0.44 (39) | -0.11 (33) |
| **Gilgit Q1-12** | -0.22 (39) | 0.01 (32) | -0.12 (39) | -0.05 (39) | 0.02 (39) | 0.00 (31) |
| **Astore Q1-12** | -0.09 (32) | 0.09 (24) | 0.01 (32) | 0.22 (32) | 0.31 (32) | 0.51 (27) |
| **Besham** | -0.02 | 0.04 | -0.12 | 0.02 | 0.15 | 0.36 |

Relationships between climatic variables and discharge from the four major dams displayed a distinct west-east trend. Tables 7-12 to 7-14 display correlation coefficients between Q1-12 and climatic variables from nearby meteorological stations. The highly ice covered Indus at Tarbela had moderate interactions with T5-9 and P11-5, not surprisingly, as its glacierised tributary basins are fed by a combination of snow and ice melt. Correlations of Q1-12 at Tarbela were intermediate between T5-9 and P11-5 (table 7-12 and 7-13), highest correlation being with T5-9 at Srinagar (r= 0.40) over a 44 year overlap and with P11-5 at Gilgit (r= 0.39) over a 48 year period. Relationships between discharge and summer air temperature decreased easterly as basin ice coverage reduced, correlations reaching a minimum in the near ice free Jhelum basin, where correlation between Q1-12 at Mangla and T5-9 at Srinagar was r= -0.39 (table 7-12). The negative relationship between Q1-12 at Mangla and T5-9 at Srinagar resulted from a slight negative relationship between precipitation and air temperature (table 7-4). Winter precipitation controlled riverflow at Mangla and Marala where westerly disturbances have a high influence on runoff, correlations being high with P11-5 at Srinagar and Srinagar district (table 7-13). Association between annual winter precipitation and annual discharge at Bhakra weakened as westerlies diminish and the monsoon intensifies (table 7-13). Annual runoff at Bhakra which has runoff influenced by the summer monsoon had high correlations with P6-9 (r= 0.55 with Shimla district). Average correlations between Srinagar district P6-9 with Q1-12 followed at Marala and Mangla, suggesting the transitional zone between winter westerlies and summer monsoon lies in the region of the Jhelum and Chenab catchments (table 7-14). Tarbela had no correlation with summer precipitation, except for a slight positive correlation with Srinagar (r= 0.23), as its glaciated northern tributaries had a negative influence on runoff and its lower, less glaciated tributaries, had positive relations with runoff.

**Table 7-9: Correlation coefficients between Q1-12 from basins in the Sutlej and T5-9. Parentheses indicate number of overlapping years between records.**

| | Shiquanhe T5-9 | Srinagar T5-9 | Jhelum T5-9 | Sialkot T5-9 | Amritsar T5-9 | Ludhiana T5-9 | Shimla T5-9 | Mukteshwar T5-9 |
|---|---|---|---|---|---|---|---|---|
| **Khab Q1-12** | 0.30 (14) | 0.02 (29) | -0.26 (31) | 0.09 (6) | 0.05 (31) | 0.66 (4) | -0.02 (16) | -0.24 (27) |
| **Rampur Q1-12** | 0.27 (22) | 0.10 (39) | -0.21 (40) | -0.10 (11) | 0.11 (42) | -0.04 (12) | 0.11 (22) | -0.29 (38) |
| **Luhri Q1-12** | 0.22 (14) | 0.14 (29) | -0.14 (31) | -0.06 (6) | 0.13 (31) | 0.14 (4) | -0.11 (16) | -0.08 (27) |
| **Suni Q1-12** | 0.33 (14) | 0.16 (29) | -0.07 (31) | -0.05 (6) | 0.12 (31) | 0.12 (4) | -0.13 (16) | -0.07 (27) |
| **Bhakra Q6-5** | 0.43 (25) | 0.00 (78) | -0.05 (54) | 0.09 (25) | 0.14 (56) | -0.13 (54) | -0.19 (65) | -0.11 (81) |

**Table 7-10: Correlation coefficients between Q1-12 from basins in the Sutlej and P11-5. Parentheses indicate number of overlapping years between records.**

| | Srinagar P11-5 | Srinagar district P11-5 | Ludhiana P11-5 | Bhakra P11-5 | Shimla P11-5 | Shimla district P11-5 | Dehradun district P11-5 |
|---|---|---|---|---|---|---|---|
| **Khab Q1-12** | 0.50 (23) | 0.36 (27) | 0.03 (10) | 0.02 (20) | 0.50 (15) | 0.38 (29) | 0.29 (29) |
| **Rampur Q1-12** | 0.36 (34) | 0.42 (36) | -0.01 (18) | 0.20 (23) | 0.26 (21) | 0.24 (38) | 0.24 (38) |
| **Luhri Q1-12** | 0.45 (23) | 0.37 (27) | -0.14 (10) | -0.08 (20) | 0.14 (15) | 0.21 (29) | 0.17 (29) |
| **Suni Q1-12** | 0.42 (23) | 0.29 (27) | -0.19 (10) | -0.05 (20) | 0.18 (15) | 0.17 (29) | 0.15 (29) |
| **Bhakra Q6-5** | 0.25 (74) | 0.19 (79) | -0.05 (62) | 0.26 (22) | 0.06 (65) | 0.02 (81) | 0.00 (81) |

**Table 7-11: Correlation coefficients between Q1-12 from basins in the Sutlej and P6-9. Parentheses indicate number of overlapping years between records.**

| | Srinagar P6-9 | Srinagar district P6-9 | Ludhiana P6-9 | Bhakra P6-9 | Shimla P6-9 | Shimla district P6-9 | Dehradun district P6-9 |
|---|---|---|---|---|---|---|---|
| **Khab Q1-12** | 0.10 (26) | 0.22 (27) | 0.12 (13) | 0.12 (20) | -0.19 (14) | -0.05 (29) | 0.13 (29) |
| **Rampur Q1-12** | 0.17 (35) | 0.27 (37) | 0.02 (21) | 0.24 (23) | -0.43 (20) | 0.18 (39) | -0.02 (39) |
| **Luhri Q1-12** | 0.24 (26) | 0.31 (27) | 0.03 (17) | 0.21 (20) | -0.03 (14) | 0.18 (29) | 0.21 (29) |
| **Suni Q1-12** | 0.24 (26) | 0.24 (27) | 0.14 (13) | 0.26 (20) | -0.04 (14) | 0.26 (29) | 0.27 (29) |
| **Bhakra Q6-5** | 0.03 (75) | 0.12 (80) | 0.17 (65) | 0.44 (22) | 0.45 (64) | 0.55 (82) | 0.42 (82) |

**Table 7-12: Correlation coefficients between Q1-12 from large dams and T5-9 from neighbouring meteorological stations. Parentheses indicate number of overlapping years between records.**

| | Gilgit T5-9 | Bunji T5-9 | Astore T5-9 | Chilas T5-9 | Hotan T5-9 | Kashi T5-9 | Shiquanhe T5-9 | Srinagar T5-9 | Jhelum T5-9 | Sialkot T5-9 | Amritsar T5-9 | Ludhiana T5-9 | Shimla T5-9 | Mukteshwar T5-9 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Tarbela Q1-12** | 0.37 (48) | 0.16 (48) | 0.25 (48) | 0.28 (46) | 0.19 (46) | 0.16 (46) | 0.35 (28) | 0.40 (44) | 0.12 (46) | | | | | |
| **Mangla Q1-12** | | | -0.23 (39) | | | | | -0.38 (72) | -0.31 (50) | | | | | |
| **Marala Q1-12** | | | 0.08 (45) | | | | | -0.05 (53) | -0.23 (54) | | | | | |
| **Bhakra Q1-12** | | | | | -0.11 | -0.06 | 0.43 (25) | 0.00 (78) | -0.05 (54) | 0.09 (25) | 0.14 (56) | -0.13 (54) | -0.19 (65) | -0.11 (81) |

**Table 7-13: Correlation coefficients between Q1-12 from large dams and P11-5 from neighbouring meteorological stations. Parentheses indicate number of overlapping years between records.**

| | Gilgit P11-5 | Skardu P11-5 | Bunji P11-5 | Astore P11-5 | Chilas P11-5 | Srinagar P11-5 | Srinagar district P11-5 | Ludhiana P11-5 | Bhakra P11-5 | Shimla P11-5 | Shimla district P11-5 | Dehradun district P11-5 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Tarbela Q1-12** | 0.39 (48) | 0.37 (37) | 0.19 (47) | 0.37 (48) | 0.31 (48) | 0.18 (38) | | | | | | |
| **Mangla Q1-12** | | | | 0.57 (46) | | 0.74 (67) | 0.42 (76) | 0.29 (59) | 0.08 (18) | 0.17 (62) | 0.27 (78) | 0.26 (78) |
| **Marala Q1-12** | | | | 0.37 (52) | | 0.53 (47) | 0.46 (52) | 0.17 (33) | 0.21 (24) | -0.05 (36) | 0.15 (54) | 0.19 (54) |
| **Bhakra Q6-5** | | | | | | 0.25 (74) | 0.19 (79) | -0.05 (62) | 0.26 (22) | 0.06 (65) | 0.02 (81) | 0.00 (81) |

**Table 7-14: Correlation coefficients between Q1-12 from large dams and P6-9 from neighbouring meteorological stations. Parentheses indicate number of overlapping years between records.**

| | Gilgit P6-9 | Skardu P6-9 | Bunji P6-9 | Astore P6-9 | Chilas P6-9 | Srinagar P6-9 | Srinagar district P6-9 | Ludhiana P6-9 | Bhakra P6-9 | Shimla P6-9 | Shimla district P6-9 | Dehradun district P6-9 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Tarbela Q1-12** | -0.07 (48) | 0.00 (37) | -0.06 (48) | -0.01 (48) | 0.07 (48) | 0.23 (39) | | | | | | |
| **Mangla Q1-12** | | | | 0.25 (47) | | 0.37 (68) | 0.37 (76) | -0.08 (62) | -0.13 (18) | -0.08 (61) | 0.01 (78) | 0.05 (78) |
| **Marala Q1-12** | | | | 0.27 (53) | | 0.43 (49) | 0.53 (53) | 0.15 (36) | 0.52 (24) | -0.09 (35) | 0.31 (55) | 0.14 (55) |
| **Bhakra Q1-12** | | | | | | 0.03 (75) | 0.12 (80) | 0.17 (65) | 0.44 (22) | 0.45 (64) | 0.55 (82) | 0.42 (82) |

# 8. Conclusion

Meteorological stations were located within close proximity to gauging stations in the UIB but were sparse at high elevations, problematic for glaciological studies as precipitation increases with elevation. Few data sets from high altitude western Himalayan climate stations led to relying mainly on district data series. There was a distinct opposing climatic regime between climate in the UIB and the Himalaya. Total summer precipitation in the UIB increased from around the 1960s to a maximum in the 1990s, while total winter precipitation increased from around the 1970s/80s to a maximum in the 1990s. Precipitation totals at Srinagar followed similar year-to-year evolutions to climate stations in the UIB but also had moderate correlations with stations in the western Himalaya. P6-9 in the vicinity of the Sutlej basin at Shimla and Dehradun decreased significantly from around the 1950s/60s but recovered in the 2000s, and had weak or negative correlations with climate stations in the UIB, suggesting sources of summer precipitation in the UIB are not of distinct monsoon origin. Negative correlations also existed between P11-5 in the UIB and western Himalaya, where P11-5 at districts Shimla and Dehradun declined by around 11% from the 1980 maxima with no recovery in the 1990s/ 2000s. Totals of seasonal precipitation were generally higher in the western Himalaya than in the UIB; this was reflected in reduced variability of P6-9 and P1-12. Variability of P11-5 had no distinct trend regionally. Variability of P11-5 reduced at Srinagar but then increased in the Himalaya as westerlies became weaker.

Year-to-year fluctuations of T5-9 in the UIB declined from their maxima in the 1970s to a minimum in the late 1980s/ early 1990s, failing to recover in the 2000s. Similar evolutions of T5-9 were followed in Tibetan climate stations; declining from the 1970s to the 1990s but recovered sharply into the 21st century. T5-9 at Srinagar had higher correlations with T5-9 in the UIB and Tibet than with stations in the Himalaya, T5-9 declining from the 1970s to the 1990s but recovering slightly in the 2000s. Mean summer temperatures at Mukteshwar increased substantially from the 1990s to 2000s and had negative correlations with T5-9 in the UIB, as did Shimla and Ludhiana. T5-9 is negatively related to seasonal precipitation. Currently the Karakoram is under a climatic phase characterised by cool wet periods, anomalous to warm dry climatic conditions in the Himalaya, and it is suggested that wet conditions may prevail over the next century in the Karakoram (Ridley et al 2013).

Riverflow from the extensively glacierised Hunza basin directly supports the Karakoram anomaly. Annual total discharge declined from the 1970s maxima to the 1990s, following a reduction of mean summer air temperature. Increasing winter precipitation since the 1980s failed to offset reduced icemelt, the melt per unit of snow being less than ice, and increasing summer precipitation retarded ablation of any exposed ice. Heightened precipitation with reduced summer temperatures and discharge suggest an overall positive mass balance of glaciers in the Hunza basin, as previously suggested (E.g. Hewitt 2005, Kaab et al 2012, Gardelle et al 2013). In the eastern Karakoram, year-to-year fluctuations of total annual riverflow at Shyok failed to support the Karakoram anomaly. Its closest meteorological station, Skardu, lies at the western end of the basin. Immerzeel (2009) imposed the eastern Shyok basin has reduced and highly variable precipitation from year-to-year. Annual riverflow mimicked T5-9 from the western Himalaya and Tibet which increased from the 1990s, therefore it is suggested glaciers in the Shyok basin are melting contrary to neutral mass gains (Bhambri et al 2012). Q1-12 at Shyok is negatively related to P6-9 and P11-5.

Inference of the state of glaciers from year-to-year annual totals of riverflows in large, less extensively glacierised basins is near impossible. Ideally discharge should be measured from pro-glacial streams of individual glaciers. Riverflows from Gilgit and Astore inclined from around the 1970s/1980s, increasingly by around 4% and 14% from the first to last decade of records, following a 13.5% and 29% rise in P11-5 from their catchment climate stations for the same period of record. It is likely that increasingly precipitation is nourishing glaciers at higher elevations although Schmidt and Nusser (2009) gave evidence for the downwasting of Raikot glacier (Astore basin) in the 21st century.

Discharge in the Sutlej basin is driven purely by precipitation. Glaciers are small and account for only a small percentage of the basin area. Annual total discharge at Khab which declined by 31% from 1972/81 to 1993/02 was negatively related to May to September mean summer air temperature but mimicked a 11% fall in winter precipitation at Shimla district for the same period. Discharge in the lower Sutlej ($Q_{ls}$), calculated from Qsuni – Qkhab, increased by around 13% for the same period as Khab (1972/81 to 1993/02), mirroring the recovery of P6-9. However, total annual discharge at gauging stations below Khab actually declined by around 15% from 1972/81 to 1993/02, increasing P6-9 insufficient enough to offset the reduction of flow at Khab.

Riverflows at Large dams had mixed long term fluctuations. Annual riverflow at Marala and Bhakra both declined following patterns of seasonal precipitation from nearby meteorological stations. Riverflow at Tarbela was stable; declining by around 4% from 1962/71 to 2000/09, while riverflow at Mangla increased by around 15% from 1960/69 to 1991/2000. Variability of annual runoff, expressed through the coefficient of variation, suggests riverflows are most vulnerable at Mangla dam as riverflows are fed directly by precipitation with little moderation by icemelt in warm dry years. Variability of Q1-12 at major dams was least at Tarbela, riverflow being moderated by runoff from its glacial tributaries in warm dry years, winter precipitation in wet winters, and summer precipitation in its lower reaches during wet summers. Statements referring to future increases in riverflow on the Indus as a result of declining glaciers (Briscoe and Qamar 2007) are not correct. Evidently, annual totals of riverflows at the outlet of the Indus at Tarbela are stable, climate in the Karakoram remains anomalous compared with surrounding areas in the Himalaya, and Tarbela will continue to moderate annual variations of runoff through moderating influences of intra-annual variations in climate. Bhakra also had low annual variability of runoff, riverflow being enhanced by the dominant monsoon precipitation from June to September and also augmented by waters from the Beas-Sutlej link canal.

Relatively low ice covered basins; Khab, Astore, and Mangla had high variability of annual runoff which reflected high annual variations of P1-12 from nearby meteorological stations. Basins which had a larger area of ice cover (Hunza, Shyok and Gilgit) had lower annual variability of runoff than low ice covered basins as icemelt compensated for reduced runoff in warm dry years, while there was still enough runoff from precipitation to compensate for runoff in cool wet years. In the UIB there was a clear concave relationship between CV of annual runoff and ice cover, while the relationship between CV of annual runoff and basin glacierisation was unorthodox and upheaval compared with previous published literature (e.g. Collins 1990, Fountain and Tangborn 1985). The true glacierisation of basins in the UIB at Tibet is unknown. Large areas, often at low elevations, are essentially hydrologically inactive, being arid and receiving little precipitation to moderate runoff. It is therefore recommended to use actual ice area in a catchment instead of percentage glacierisation of a basin.

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